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DYNAMICS OF A CHERNOZEMIC SOIL SYSTEM

by

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ABSTRACT

A Gleyed Eluviated Black Chernozemic pedon in the Lake Edmonton plain of central Alberta was investigated to determine if its salient properties reflected the character of the pedogenic processes operating in the present environment. Unlike for other parts of the Black soil zone, historical, pollen, and phytolith evidence suggest a dominantly forest vegetation at the site during the late Holocene, with the present Populus balsamifera forest being typical of the Boreal Transition zone.

The pedogenic influence of the present ecosystem is expressed in several aspects which seem compatible with the existence of a well-structured, base-saturated Chernozemic A horizon. First, the litterfall returns a considerable quantity of bases to the soil, with the dominant understorey shrub, Cornus stolonifera, being important in calcium cycling. Second, soil faunal activity appears significant in fabric development within the organic and Ah horizons. However, soluble organic substances derived from leaching of the forest canopy and litter materials, as well as the effects of physical processes, may play a role. Third, the large transpiration demands of the forest vegetation rapidly deplete soil moisture reserves during the growing season, thereby reducing the potential for leaching.

A high degree of clay-organic matter complexation in the Ah horizons and a lack of major clay mineral alteration are significant pedon characteristics shared with other Black Chernozemic soils.

The present groundwater regime at the site is characterized by discharge which does not appear to affect the control section. However, a temporary perched water table can occur during the spring snowmelt, but

it may not be sufficient to produce the weak hydromorphic features in the pedon. Drainage conditions have likely improved since settlement, so these features may be relict.

Classification of this pedon as a Chernozem is not consistent with the original usage of the term to designate grassland soils lacking B horizons or incipient eluviation. A more suitable grouping drawn from European pedology is with the Gray Forest soils. These are treated as a distinct ecological soil type of the steppe-Boreal Forest transition, possessing both a prominent Ah horizon and some degree of eluviation. It may also be fruitful to recognize the hydromorphic genetic pathway, that of the Meadow Chernozems, as being applicable to some of the Black soils in western Canada.

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CHAPTER 1

INTRODUCTION

An earlier study of Orthic Black Chernozemic soils (Dudas, 1968) found that regional variations in Ah horizon structure, apparently related to clay-organic complex properties, existed within the Black soil zone in Alberta. Furthermore, some pedologists have observed that the soils developed on clay-rich glaciolacustrine parent materials in the northern part of that zone possess certain distinctive Ah horizon properties: a more grayish colour, perhaps associated with impaired drainage, and a stronger granular structure (T.W. Peters, Agriculture Canada, personal communication). Since the region of Black Chernozemic soils encompasses more than one vegetation zone in Alberta, the relationship between the genesis of these soils and environmental factors is evidently complex, yet has received insufficient attention.

In view of these observations, this study was initiated to meet the following central objective: to determine if the Black Chernozemic soils of the Edmonton area could have formed by processes operating in the same fashion as in the present environment. In order to accomplish this purpose, two subsidiary objectives were identified: (1) to relate the properties of the pedon to both observed and inferred processes in the soil system, and (2) to reconstruct the environmental changes during the Holocene that may have controlled the direction of pedogenesis in the region.

The organization and content of this thesis can be outlined as follows. Chapter Two presents the background to the central objective by reviewing the literature on three related issues: concepts of

Chernozem genesis and classification, soil genesis in relation to the forest-grassland transition, and the formation of mull A horizons in diverse environments. Chapter Three briefly discusses the rationale for this study's dual emphasis on both inferred processes and observed soil dynamics, and presents an account of the methods used. Chapter Four reviews the aspects of Holocene environmental change relevant to the study: glacial history, climatic fluctuations, and changes in vegetation zonation. The main body of data and its interpretation are contained in Chapter Five which examines the results of both field and laboratory studies of pedon properties and genetic processes. The concluding chapter brings together the principal findings of this study in order to assess the degree to which the central objective was met and to recommend future directions for research.

CHAPTER 2

BACKGROUND TO THE PROBLEM

2.1 Introduction

As was pointed out in the previous chapter, regional variations in properties of Chernozemic A horizons have been observed within the Black soil zone in Alberta. Such differences suggest that there may be a corresponding variation in pedogenic factors and processes within the zone. Since the placing of soil zone boundaries depends in part on the taxonomic system used and its underlying genetic concepts, it is important to appreciate the background of such systems and concepts. Hence, this chapter will review the varying concepts of Chernozemic soils and their genesis, both in Canada and abroad. Because of the important role attributed to vegetation in Chernozem genesis and transformation, the question of forest-grassland transitions in relation to soils will be examined. Finally, the more general issue of mull Ah horizon formation in diverse environments will be considered.

2.2 Chernozemic Soils: Classification and Genesis

The Chernozem occupies a special place in the history of pedology, since it was the subject of Dokuchaev's pioneering investigations of soil geography (Joffe, 1949; Buol, Hole, and McCracken, 1973). Over the expanse of the Russian steppes, Dokuchaev observed the correspondence of climatic gradients and vegetation zones with the pattern of soil types and formulated his factorial model of soil genesis. From this beginning, classification and mapping of Chernozemic soils has extended to other continents, often with differences appearing in the taxonomic and genetic

concepts of various workers.

As discussed in pedology texts, the Chernozem is a soil of the mid-latitude grasslands, developed in a continental climate of low rainfall and humidity with hot summers, cold winters, and limited supplies of moisture available for leaching. The original concept of the Chernozem, literally "Black Earth", involved an accumulation of humus in a prominent Ah horizon possessing a definite granular structure. Other characteristics include the lack of significant mineral weathering or redistribution of clay and a neutral to weakly alkaline reaction caused by carbonate accumulation in the profile (Gerasimov and Glazovskaya, 1965; Joffe, 1949).

In Russian pedology, the Typical Chernozem has an A-C profile as its central concept. There may be a transitional horizon between the A and C horizons but it is calcareous. Further subtypes of the Chernozems are designated according to the depth of carbonate leaching, A horizon thickness, and organic matter content (Gerasimov and Glazovskaya, 1965). The Ordinary and Southern Chernozem subtypes have lower humus contents, less leached profiles and a secondary carbonate horizon closer to the surface; the latter subtype may have gypsum at depth. Calcareous and Residually Calcareous Chernozems contain carbonates to the surface, reflecting the high lime content of their parent materials. Solonetzic Chernozems occur on saline parent materials and share some properties of Solonetzic soils. The Leached Chernozem subtype occurs in the forest-steppe zone and is identified by its non-calcareous B horizon which may be slightly enriched in clay. The Podzolized Chernozems differ in degree from the previous subtype, having a stronger clay accumulation in the B horizon and signs of eluviation at the base of the A horizon (Ivanova

and Rozov, 1958). The Meadow Chernozems are recognized as a distinct type and occur in poorly drained sites characterized by gleying and salinization.

European pedologists have maintained the A-C profile as the central concept of the Chernozem. For example, Kubiena (1953) defined the Typical Chernozem in terms similar to those of the Russian pedologists. In his classification, the A-B-C profiles of the steppe were referred to as Degraded Chernozems. Duchaufour (1977) distinguished the leached Chernozem from the typical Chernozem on a similar basis; the latter has only an incipient blocky B horizon.

In North America, the term Chernozem has been adopted, but with a somewhat differnt usage. Recent Canadian systems of soil classification have defined a Chernozemic Order which includes lighter coloured steppe soils (Brown and Dark Brown) as well as Black Chernozems. In Russian and European classifications, the Brown and Dark Brown (Chestnut) soils are separated from the Chernozems as distinct types. A second major departure in the Canadian classification is in the central concepts of the various Chernozemic Great Groups. The Orthic subgroup has an A-B-C horizon sequence, whereas the nearest equivalent of the Russian Typical Chernozem is merely another subgroup, the Rego Black (Canada Soil Survey Committee, 1978). The current U.S. Soil Taxonomy (Soil Survey Staff, 1975) resembles the Canadian system in that Chernozemic soils (in the Russian sense) are grouped with other soils having a dark, organic matter-enriched surface horizon, the Mollisols. This order also includes the group formerly known as the Prairie soils, similar in morphology to the Russian Leached Chernozems but lacking a horizon of secondary carbonate accumulation.

The Chernozemic order of the Canadian soil taxonomic system includes other soils which in Russian and European pedology are designated as distinct genetic types. In particular, the Eluviated Black and many of the Dark Gray Chernozems may correspond to those referred to as Eluviated Danubian Chernozems and Gray Forest Soils, respectively, by Duchaufour (1978), or as Dark Gray and Gray Forest Soils in Russian terminology (Gerasimov and Glazovskaya, 1965).

A description provided by Parfenova et al. (1964) of a Gray Forest Soil in the deciduous forest zone of the U.S.S.R. suggests many similarities with the Dark Gray Chernozemic soils: a thick Ah (and possibly Ahe) horizon (0-35 cm), a much thinner and bleached eluvial horizon (35-45 cm) and a blocky illuvial B horizon. Thus, despite the similar name, the Gray Forest Soils are not the same as the Gray Luvisols (formerly Gray Wooded soils) of Canada. In the latter, the Ah horizon is thin or absent (Moss and St. Arnaud, 1955). Fridland and Erokhina (1976) appeared reluctant to consider the black soils of the Aspen Parkland as Chernozems. Instead, only the thinner black soils of lower humus content and the dark brown soils were correlated with the Chernozems of the U.S.S.R. Apparently, the thicker northern black soils were viewed by these Russian pedologists as being of meadow or hydromorphic origin. In the absence from the Russian literature of explicitly stated taxonomic criteria based on soil properties, any of these correlations are necessarily tentative.

To conclude this discussion of classification, it is emphasized that the focus of this thesis is on the Black Chernozemic soils as defined in the Canadian soil taxonomic system. The A-B-C horizon sequence which constitutes the central concept of these soils, has its equivalents in the Degraded or Leached Chernozems. In North America,

these soils occur principally in the eastern Great Plains region in a broad belt extending from Texas northwards through the Dakotas, continuing through southeastern Manitoba, central Saskatchewan and Alberta and narrowing to a band along the edge of the Rocky Mountain Foothills (Soil Survey Staff, 1975; Clayton et al., 1977).

The salient properties of the Chernozem, as it has been defined in North America, can be listed as follows (Dudas and Pawluk, 1969b; McClelland et al., 1959; Nygard et al., 1959; Redmond and Omodt, 1967; St. Arnaud and Whiteside, 1963):

(1) absence of L-H horizons,

(2) a black Ah of granular structure, rich in organic matter of the humic acid type which tends to be intimately associated with mineral material,

(3) approximately neutral pH values and a high degree of base saturation, especially by calcium, which is thought to maintain the soil colloids in an immobile, flocculated condition,

(4) B horizons, which may appear illuviated in the field, but not always in thin section, characterized by oxidation, carbonate removal, and weak or moderate prismatic macrostructure,

(5) secondary carbonate accumulation at depth, often as a distinct Cca horizon, and

(6) generally mild weathering of primary minerals, with no major transformations of clay minerals.

To account for these properties, genetic concepts of Chernozemic soils have emphasized the role of vegetation and biological activity

in contributing and distributing organic matter. Of equal importance is the effect of climate in governing rates of organic matter decomposition and transformation, and through the soil moisture regime, in controlling the transport of soluble constituents.

Both past and contemporary Russian pedologists have emphasized the crucial role of grassland vegetation in Chernozem formation (Joffe, 1949; Gerasimov and Glazovskaya, 1965). A thick cover of grass and forbs, together with a dense network of roots, provides a large annual addition of nutrient-rich, readily decomposed organic matter. The root systems also act to take up available nutrients and prevent their removal from the solum, while helping to aggregate the soil through their growth pattern and decomposition products.

Since calcium is present as carbonates in the Chernozem parent material and is cycled through the vegetation at a considerable rate, base saturation is high. Kononova (1975) and other writers attributed the granular structure and stability of Chernozem humus to the flocculating effect of this abundant calcium. A major role in structure formation is also played by soil fauna, particularly earthworms (Gerasimov and Glazovskaya, 1965).

The climatic influence on Chernozem formation is twofold. A generally non-flushing moisture regime limits the degree of leaching, while the occurrence of wetting and drying cycles is thought to assist in the polymerization of humic substances (Kononova, 1975).

It is noteworthy that the Russian pedologists have recognized another soil type, the Meadow Chernozem, which has a distinct genetic pathway. Occurring in sites with impaired drainage, these soils have A horizons similar to those in other Chernozems, but show gley features,

as well as some degree of salinity (Gerasimov and Glazovskaya, 1965). This second genetic pathway does not seem to have received much attention from North American pedologists, even though some Chernozemic soils may have evolved from Humic Gleysols.

2.3 Soils and the Forest-Grassland Transition

As the previous section has shown, Chernozems are regarded as being grassland soils, so when their genesis is contrasted to that of the Luvisols, the differences are attributed to the role of vegetation. Under grassland, organic matter is added both at the surface and by root death, while forest vegetation contributes mostly leaf litter. In the latter case, more rapid decomposition is thought to produce mobile, low molecular weight organic acids which can assist colloid translocation through the profile, giving rise to Luvisolic soils (Pettapiece, 1969). The coincidence of the grassland-forest ecotone with the transition from Chernozemic to Luvisolic soils in western Canada is cited as evidence for such a vegetation control of soil genesis.

This example from western Canada illustrates what Jenny (1941) called a biosequence, a series of soils formed along a gradient of vegetation change. Numerous studies in the U.S. midwest have examined such sequences along the transition from tall-grass prairie to deciduous forest (Bailey et al., 1964; Geis et al., 1970; Severson and Arneman, 1973; Smith et al., 1950; White and Riecken, 1955). The corresponding soil transition is from Brunizems (Prairie soils) to Gray-Brown Podzolics, with the principal variations in soil properties being:

(1) thinning of Ah horizons and less organic matter accumulation at depth in the profile,

(2) lowering of pH and base saturation, especially in the B horizon,

(3) greater depth of carbonate removal,

(4) development of an eluvial horizon,

(5) clay illuviation in the B horizon, and

(6) increased mineral weathering.

When vegetation change occurs, for example, forest invasion of grassland, such sequences are used as models of the transformations assumed to occur in soil properties. What are not clearly established, however, are the rates and controls of such changes. In western Canada, Pettapiece (1969) cited anecdotal evidence of visible eluviation of a Chernozem occurring in about 50 years after planting of poplars. On a similar time scale, Dormaar (1967) found evidence for changes in Chernozem humic acid infrared spectra which were attributed to less than 60 years of occupancy by *Pinus sylvestris*. Similarly, Dormaar and Lutwick (1966) examined Chernozemic soils in a region of southwestern Alberta in which poplar encroachment on grassland was inferred from historical records. After approximately 80 years, changes were detected in profile morphology, chemical properties (reduction in pH and base saturation), and particularly in humic acid infrared spectra.

However, other evidence suggests that the rate and direction of soil transformation after forest invasion of grassland may be more variable. At an Illinois site forested for an estimated 400 to 600 years, Geis *et al*. (1970) found marked differences in the rate of transformation of Prairie soils toward Gray-Brown Podzolics. Vegetation and parent material were fairly homogeneous so the only variable appeared to be drainage, as controlled by topography. Partially transformed soils occurred on sites having higher relief and with water tables

below the solum throughout the year. However, Prairie soils persisted at lower sites at which water tables remained in the upper solum until late spring. The authors concluded that the better internal drainage at upland sites assisted the leaching processes responsible for transforming Prairie soils. They also suggested that the unaltered Prairie soils formed a continuum with the Humic Gley soils occupying the lowest topographic positions. In view of these findings,it is significant that the other studies in the midwest region cited previously all examined biosequences on well-drained sites.

A similar situation in the Canadian Prairies was described by Pettapiece (1969). He noted that in some areas of Black and Dark Gray Chernozemic soils, drainage conditions have likely improved since settlement. These soils probably developed under meadow or balsam poplar-willow vegetation associated with moister conditions, so their relationship to the well drained soil sequence of the forest-grassland transition was not clear. This and the previous reference appear to be the only discussions in the North American literature which suggest a genetic pathway at all similar to that of the Russian Meadow Chernozems.

A final example, which also complicates the picture of Chernozem genesis in relation to vegetation type, comes from Russia (Afanas'yeva, 1966). Undisturbed profiles of Chernozems with thick Ah horizons were studied under both steppe and oak forest. Apart from a more pronounced granular structure and a greater depth to the peak secondary carbonate accumulation under the latter vegetation type, the properties and morphology of the two profiles were similar. Comparing water regimes, the author found a greater fluctuation in soil water content under forest, with higher snow accumulations, but greater soil moisture with-

drawal for evapotranspiration, resulting in only occasional wetting of the entire profile. Although the potential for leaching of carbonates was greater in the forest soil, this was apparently compensated by a fivefold greater return of calcium through litterfall. Afanas'yeva's conclusion was that where moisture conditions permitted forest to occur in the steppe region, it did not seem impossible for this vegetation to coexist with Chernozemic soils.

2.4 Mull A Horizons in Diverse Environments

As discussed by Duchaufour (1976), mull is characterized by high biological activity, rapid litter decomposition and good crumb structure resulting from intimate combination of mineral and humic components. Despite these properties in common, many types of mull exist, characterized by different properties and modes of evolution. It seems clear from a reading of the literature from western Canada that there is often an implicit assumption that the mull Ah horizon is a feature incompatible with genesis in a forest environment. Studies already cited have shown that the evidence supporting this belief is at best equivocal. Although the entire subject of mull Ah horizons is a large one, this review has the limited objective of showing that this horizon type is a polygenetic feature with several agents and processes active in its formation.

While the Ah horizon, as defined in Canada (Canada Soil Survey Committee, 1978) has very minimal requirements*, the Chernozemic A horizon has several very specific criteria (Appendix). Despite this specificity, the authors of the current Canadian soil classification

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* The Ah horizon colour value is $\geq$ 1 unit lower than that of the underlying horizon and/or it has 0.5% more organic carbon than the IC horizon. Organic carbon content is 17%.

system were obliged to introduce a climatic requirement (mean annual soil temperature $>0^{\circ}C$, soil moisture class drier than humid) into the definition of a Chernozemic A horizon. This climatic restriction is apparently all that prevents the Ah horizons present in the Gray-Brown Luvisols and Melanic Brunisols from being designated as Chernozemic. Even with this restriction, some uncertainty still exists regarding the status of some Melanic Brunisols, particularly in subalpine and alpine areas.

Reference was made earlier to the Gray Forest Soils of European and Russian soil classifications, which appeared to share certain characteristics with some of the soils designated as Chernozemic in Canada. Judging from the descriptions provided, the Ah horizons in the former could also be characterized as being of the mull type and seem to resemble those of the Gray-Brown Podzolic (Luvisolic) soils (Gillespie and Elrick, 1968; Rutherford, 1967; Stobbe, 1952).

Much attention has been given to the role of faunal activity in the formation of the mull humus form. Russell (1973) emphasized the physical and structural effects of soil macrofauna -- those animals large enough to enlarge pore sizes by their movements. Apart from the improved soil aeration resulting from faunal channels, the major effects of these organisms are through their ingestion and excretion of litter material and its incorporation in the mineral soil. These faunal excretions can comprise the mull humus form, particularly in the case of earthworms which ingest soil along with plant debris. The stability of earthworm casts seems related to factors such as length of fungal hyphae and content of polysaccharide gums. In addition, the secretion of mucoproteins as an external film on earthworm bodies may

help to preserve the pore spaces created by their movements through the soil.

The Gray-Brown Luvisols, along with some Melanic Brunisols (formerly called Brown Forest Soils) are typical of a temperate forest environment and show a marked influence by soil macrofauna. As shown by Neilson and Hole (1964), earthworms in a Gray-Brown Podzolic soil were able to incorporate virtually all of the annual litterfall and were estimated to be capable of forming a mull Ah horizon in 30 to 40 years. It seems widely accepted that such activity is characteristic of only base-rich, near-neutral pH soils, yet Langmaid (1964) found that where artificially introduced, earthworms thrived in strongly acidic Podzols under coniferous forest in New Brunswick. This study showed that in a very short period of time, three years or less, such invasions could produce an "Ap" horizon from complete incorporation of the organic horizons in the mineral soil. This worm-worked horizon was described as having the strong granular structure which is usually typical of mull.

Although it is uncertain whether or not the role of the soil fauna is of equal importance in the genesis of all mull Ah horizons, recent work has increased our understanding of the chemical and biochemical factors involved. Field and laboratory studies by Duchaufour and his students have pointed out the roles played by iron, clay, manganese, and calcium in mull formation. Indeed, some of their findings indicate that the influence of these agents in reorganizing soil material can sometimes outweigh the effects conventionally attributed to particular types of vegetation litter.

Duchaufour and Souchier (1978) found an inverse relationship between parent material Fe content and the degree of podzolization in acidic soils. Mull was associated with the parent materials of highest Fe content. In subsequent laboratory incubation experiments, iron hydroxide added to decomposing litter increased the yield of insoluble humin. This result was attributed to direct acceleration of litter decomposition and/or an indirect effect through conversion of organic substances to less soluble forms.

In addition to Fe, Mn also appears to have an important indirect role in the humification process. Duchaufour and Jacquin (1974) and Duchaufour (1973) noted that although Mn is not abundant in most parent materials, it is concentrated in mull humus by biocycling, much more so than in other humus forms. Unlike in the case of Fe, the inherited parent material content of Mn is of less importance in controlling its abundance in the biologically active part of the soil; however, both elements seem to have similar roles in humification. The authors speculated that Mn is contained in a cofactor of polyphenoloxidase which is involved in the polymerization of stable humic substances.

A certain minimum silicate clay content was found to be necessary for mull to be produced in experimental soil parent material columns incubated with litter materials (Vedy and Jacquin, 1972). While mull did not form under a slowly decomposable, low nitrogen litter, regardless of clay content, an easily biodegradable, nitrogen-rich litter produced mull only when 5 to 6% silicate clay was present. The same study also supported the importance of Fe in the humification process and suggested that stable clay-iron-humus complexes are a distinctive feature of mull in acidic soils. In the somewhat different environment

of the Chernozems, Duchaufour (1976) attributed the stability of their organic matter to the effect of swelling clays in catalyzing the polymerization of humic acids, although the mechanism responsible was not elaborated on.

Much of the traditional thinking about Chernozemic soils has viewed their high base status, especially the level of exchangeable calcium, as somehow causally related to the stability of their organic-mineral complexes. Only recently has the function of calcium in the formation of these complexes been studied in detail. Duchaufour (1973) cited evidence which demonstrated that calcium in mull tends to concentrate in the non-extractable humin fraction. Unlike Fe-humus complexes which can be broken down by decomplexing agents, these associations of calcium-rich humin and clay are much more stable. Calcium originating in the litter layers is initially soluble and can be transported through the soil, but becomes bound up in these stable complexes whose non-extractability seems related to their calcium content (Duchaufour and Jacquin, 1974).

Although many of the foregoing observations concern acidic soils of more temperate climates, it does seem worth considering that Chernozems may share some of the same stabilizing mechanisms involved in mull formation in other environments.

2.5 Summary

The key points from this review of the background issues in this study are as follows.

(i) Important differences exist between the central concepts of Chernozemic soils as defined by European and Russian pedolo-

gists as against those developed by their North American counterparts.

(2) More than one genetic pathway for Chernozems is recognized by Russian pedologists, while one of these, the hydromorphic sequence involved in the formation of Meadow Chernozems, seems only to have been hinted at in the North American literature.

(3) Some of the soils included in the Chernozemic Order in Canada are recognized as a separate genetic type (Gray and Dark Gray Forest Soils) by Russian pedologists.

(4) A regular transition between Chernozemic and Luvisolic soils, correlated with that between grassland and forest vegetation, has been observed in western Canada. As in the analogous biosequence in the U.S. midwest between the Prairie and Gray-Brown Podzolic soils, attention has been concentrated on well drained sites. However, anomalies related to drainage conditions, as well as uncertainties in the postulated rates of soil transformation after vegetation change, suggest that this sequence should be examined more critically.

(5) Mull Ah horizons are not unique to the Chernozems. This is not a new observation, but it appears that only recently has attention been paid to the variety of factors involved in mull formation. While mesofaunal activity is recognized as a factor in mull formation, at least in some soils, certain other chemical and mineral agents also play a role. Uncertainties as to the genesis of mull are illustrated by the fact that it can occur under conditions (acidic soil reaction, forest vegetation) which

are often assumed to be responsible for Chernozem degradation. Perhaps the emphasis on high base saturation as a requirement of the Chernozemic type of mull has directed attention away from more genetically significant properties. Therefore, it seems reasonable to suggest that our understanding of the mechanisms of alleged Chernozem "degradation" under forest vegetation is somewhat deficient.

CHAPTER 3

METHODS AND MATERIALS

3.1 Introduction

Most of this chapter is devoted to an account of the field and laboratory techniques used in this study. First, however, it is necessary to explain the rationale for the general approach of this study, which is to complement the normal techniques of pedon characterization by studies of processes occurring in the soil system and by examination of relevant aspects of environmental history.

Traditionally, pedologists have carried out their studies of soil genesis in two main ways. The most common is to observe and sample a pedon or pedons in the field and conduct a variety of analyses in the laboratory. From these data, generalizations are made about pedon properties and their inter-relationships. These observations can then be used to test or suggest hypotheses about how such properties came to be. Another approach is slightly different and begins with the selection of a sequence of pedons that accompanies some systematic variation in a property of the environment or simply the time elapsed since pedogenesis began. This is the familiar sequence method: biosequences, climosequences, and chronosequences (Jenny, 1941). Correlations are then sought between variations in the properties of the soils and variations in the other factors: organisms, climate, time, and so on.

From both of these approaches have come the core of the discipline of soil genesis and classification. These results have included large amounts of data, several useful generalizations, and a certain

amount of dogma.

Both of these methods are subject to the problem of equifinality, that is, similar results arising from different causes or processes. As pointed out by Chorley and Kennedy (1971), equifinality poses serious difficulties for attempts at explanation in the earth sciences. A given condition of a system can result from processes that may be simulated by any number of plausible models. Hence, inferences made about genesis from the properties of a system as observed at one moment in its history are only a prelude to explanation.

In other parts of soil science, for example, those related to soil management in agriculture and forestry, two other approaches, which are often combined, have become important if not dominant. These involve (1) controlled manipulations of field conditions or laboratory microcosms and (2) fine-scale observations of variations over time in a restricted set of related soil properties. These experimental methods have been finding a more important place in soil genesis studies (e.g. Hallsworth and Crawford, 1965) for a very important reason. Taking this study as an example, if we want to understand how a barren deglaciated surface became the Black Chernozemic soil that we observe 10,000 years later, it is essential to know what is happening in the soil system and why.

The emphasis in soil genesis studies is therefore becoming twofold: processes as inferred from properties and as observed in field or laboratory systems. Since the term "process" is used rather freely by pedologists, it requires definition. For the purposes of this study, the definition is a pragmatic one; a process is something that is observed to happen in a given system. No particular scale or level

of abstraction is implied, since identification of a process for study depends both on what is already known and the objectives of the investigation.

Finally, since a subsidiary objective of this study is to examine the aspects of Holocene environmental history pertinent to the central purpose, it is important to realize that the soil has a memory. In other words, some processes leave evidence that persists long after the conditions that produced it have changed or ceased to exist. Hence, the historical dimension in pedology must always be appreciated so as to complement or correct genetic explanations based on current properties and processes.

3.2 Site Selection

The site chosen for this study is located at the Ellerslie Research Station of the University of Alberta (N.E. ¼, Sec. 24, Tp. 51, R. 25, W.4), about 12.5 km south of the main campus and at an elevation of 686 m. The pedon chosen for characterization and process monitoring occurred under a 50 m wide strip of Balsam Poplar (*Populus balsamifera*) forest along the western boundary of the Station. Although the soils associated with this stand appeared to be moderately well to imperfectly drained, which is not part of the central concept of the dominant soil series mapped in the area (Malmo SiCL), several factors favoured selection of this site: (1) the soil profile showed no evidence of disturbance by ploughing, (2) the vegetation appeared to be typical of that described in the Edmonton area prior to settlement, and (3) its location on supervised University property reduced the risk of vandalism to the proposed installations.

3.3 Analyses Performed on Whole Soil Samples

Routine physical and chemical analyses to characterize the pedon were performed on air dried samples (passed through a 2 mm sieve) according to the methods outlined by the Canada Soil Survey Committee (McKeague, 1978), unless indicated otherwise.

Ten gram samples for particle size analysis were pretreated with H_2O_2 for organic matter removal and buffered Na-acetate (pH 5) for carbonate removal. The samples were then dispersed by ultrasonification (3 minutes at 400 watts output, Braunsonic 1510 Ultrasonic Probe) and addition of sodium hexametaphosphate, followed by wet sieving (50 μm). The sand was weighed and stored and the remaining silt and clay suspension analysed by the pipette method. Fine clay (< 0.2 μm) was determined by centrifugation.

Soil reaction was measured in a saturated soil-distilled water paste using a Corning Model 12 Research pH Meter equipped with Corning pH and reference electrodes.

Exchangeable cations were displaced by 1 N buffered ammonium acetate (pH 7), with total cation exchange capacity determined by steam distillation of NH_3 from ammonium-saturated samples. C.E.C. was also estimated by summing the quantities of exchangeable cations. Contents of Ca^{++}, Mg^{++}, K^+, and Na^+ in the extracts were determined by atomic absorption spectrophotometry (A.A.S.). The instrument used for this and all subsequent A.A.S. procedures was the Perkin-Elmer 503 model. To suppress interferences, all A.A.S. determinations of Ca^{++} and Mg^{++} in this study were performed on dilutions and standards containing 2000 mg l^{-1} La^{+++}.

The Walkley-Black method for organic carbon analysis was used and total nitrogen was measured by the semi-micro Kjeldahl method.

Calcium carbonate equivalent was determined by the method of Bundy and Bremner (1972).

Extraction of iron and aluminum oxides was carried out by three techniques, with determination of Fe and Al concentration in the extracts performed by A.A.S. The sodium pyrophosphate extraction removes organic-complexed Fe and Al (McKeague, 1967). The acid ammonium oxalate method extracts both organic-complexed and amorphous Fe oxides (McKeague and Day, 1966), while the dithionite-citrate-bicarbonate technique (Mehra and Jackson, 1960) is taken as a measure of the total pedogenic Fe oxides. The interpretation of the latter two methods in terms of differentiation of forms of Al is less clear (McKeague et al., 1971).

Humic acids from the L-F, H, Ah1, Aegj and Bmgj horizons were extracted by the Na-pyrophosphate and NaOH procedure employed by Dormaar (1967). Extracted humic acids were dialysed against H^+ -resin in deionized water which was changed once or twice daily over a one week period. The humic acids were pressed into KBr pellets for which infrared spectra were obtained (Beckman IR-20 infrared spectrophotometer).

Micromorphological studies were carried out on 7.5 x 5 cm thin sections cut from intact sample blocks impregnated with Scotchcast epoxy resin.

3.4 Analyses Performed on Soil Particle Size Separates

Particle size separations were performed on 50 to 100 g samples which were dispersed ultrasonically, as described in the previous

section, prior to wet sieving (50 μm). The 50-250 μm fractions were sieved from H_2O_2 - treated and carbonate-free samples and separated into density fractions by centrifugation in bromoform (sp gr 2.92) following procedures described by Carver (1971). The < 2.92 sp gr separate should contain all of the feldspars, while excluding other Na- and Ca-bearing minerals. Sub-samples of this density fraction were dissolved by HCl-HF treatment (Pawluk, 1969) and Ca^{++} and Na^{+} were determined by A.A.S. The soda-calcic feldspar content was estimated by assigning all Na and Ca to this mineral group.

For the study of opal phytoliths, H_2O_2 - treated Ah horizon samples from four sites in Alberta were fractionated by sedimentation after wet sieving for sand removal. Clay was removed and the remaining silts were fractionated into fine (2-5 μm), medium (5-20 μm), and coarse (20-50 μm) fractions. The opal phytolith content of the medium and coarse silts was determined by a method similar to that used by Norgren (1973). Approximately 400 grains were counted in each sample under a petrographic microscope and the proportion of opal present was multiplied by the percentage content of the size fraction in the total soil, giving an estimate of the opal content of the soil horizon. All three fractions from the Ellerslie Ah1 sample were subjected to a density separation using centrifugation in a tetrabromoethane and bromobenzene mixture (sp gr 2.30). The < 2.30 sp gr fraction was examined for phytolith morphology under a Cambridge Stereoscan S4 scanning electron microscope. Samples were coated with 150 Å of Au in an Edwards Sputter Coater.

Ca-saturated clay fractions were separated into fine (< 0.2 μm) and coarse (0.2-2.0 μm) fractions by centrifugation, freeze-dried and

subjected to the following analyses.

(1) X-ray diffraction analysis was performed with a Phillips diffractometer and Cu-K-α radiation, with seven sample pretreatments:

(a) Ca-saturated, 54% relative humidity,

(b) Ca-saturated, glycerol solvated,

(c) Ca-saturated, ethylene glycol solvated,

(d) K-saturated, $105^{o}C$, 0% relative humidity

(e) K-saturated, 54% relative humidity,

(f) K-saturated, $300^{o}C$, and

(g) K-saturated, $550^{o}C$.

Clay samples were mounted on glass slides by the paste method (Theisen and Harward, 1962).

(2) C.E.C. was determined on Ca-saturated, H_2O_2 - treated samples by extraction with 2 N NaCl and measurement of the displaced Ca^{++} by A.A.S.

(3) Surface area was measured by the ethylene glycol monoethyl ether method (Carter et al., 1965), with the modification that samples were not treated with liquid EGME but were equilibrated with its vapour.

(4) Mica content was estimated from the K_2O content of an HCl-HF dissolution analysis, assuming a 10% K_2O content in the mineral.

(5) Clay-organic complexation in Ah1, Ah2, and Ah3 horizon clay fractions was studied through comparison of total and pH - dependent cation exchange capacities determined on samples with and without H_2O_2 treatment (Dudas and Pawluk, 1969a).

(6) Total carbon content of the untreated clay fractions used in (5) was determined using the LECO induction furnace.

3.5 Field Studies

In spring, 1976, the following equipment was installed within or beside a fenced enclosure at the site chosen for the study:

(1) a nest of three piezometers with depths of 4.2, 7.3, and 10.4 m and a water table well (3.6 m),

(2) Soil Test Fiberglass soil moisture cells (MC3IDA) in the H, Ah, Bmgj, Ckgj1, and Ckgj2 horizons, and

(3) a Soil Test Thermistor Temperature Probe (G160).

Monitoring of this equipment was carried out weekly during April-November, 1978, May-October, 1979, and April-May, 1980. During the intervening winter periods, temperature readings were recorded at three or four week intervals.

Three lysimeters were installed adjacent to the enclosure in May, 1978. Using hydraulic pressure from a coring truck, the plastic cylinders, equipped with a screw-mounted cutting edge, were forced into the soil until their tops were at the same level as the surface. The cylinders and their enclosed soil columns were then removed and an appropriate thickness of soil was taken from the bottom of the column. The remaining soil in the columns comprised the horizon sequences (1) L-F, (2) L-F, H, and (3) L-F, H, Ah1, Ah2. In the space remaining in the cylinders below the soil columns, the collection apparatus was installed, consisting of a 125 ml plastic beaker fastened below a funnel filled with washed plastic beads. The leachate sampling tube extended from the beaker, passed through a hole in the side of the cylinder and was attached to the en-

closure fence. A hand pump was used to collect the leachates and the lysimeters were checked with the same frequency as the other equipment during the growing season.

Precipitation was sampled in an open field 150 m east of the enclosure, using two collectors each consisting of a 20 cm diameter plastic funnel, with a 2 mm wire screen at its base, emptying into a 5 l plastic jug. Similar collectors were used to sample throughfall under the forest canopy near the enclosure. Collector locations for the latter were selected to include three typical vegetation cover types: (1) a high *Populus balsamifera* canopy with a 5-10 m layer of *P. tremuloides*, (2) a high (15-20 m) *P. balsamifera* canopy, and (3) a *P. balsamifera* canopy with a dense 1-2 m layer of *Cornus stolonifera* shrubs. Stemflow was collected from two *P. balsamifera* trees with the collection apparatus consisting of an inclined rubber gutter secured to the trunk by epoxy and nails, emptying via a funnel and tubing into a 5 l plastic jug. Sample collectors for precipitation, throughfall, and stemflow were changed weekly during the growing seasons of 1978 and 1979.

Water samples from these four sources (precipitation, throughfall, stemflow, and lysimeter leachate) were analyzed for a variety of inorganic soluble constituents. Samples were often of insufficient volume to permit the full suite of tests to be performed. This was particularly the case for the lysimeter leachates, so some of the analytical methods were modified to reduce the sample volumes required.

After collection, all samples were filtered with Whatman #41 ashless filter paper, although discrete particulate debris was usually not observed in the lysimeter leachates. After filtration, pH and bicarbonate were determined as soon as possible, after which samples were frozen until

other analyses could be performed. Upon thawing, these filtered samples did not show evidence of precipitation of previously soluble constituents.

Inorganics measured consisted of:

(1) pH, determined with a Beckman Electromate pH meter equipped with a Fisher MicroProbe combination pH electrode,

(2) bicarbonate, measured by titration to pH 4 with standardized 0.01 N H_2SO_4,

(3) sulphate, by a turbidimetric procedure using SulfaVer IV pre-measured reagent pillows (Hach Chemical Co., Ames, Iowa) and a conditioning reagent, with turbidity measured by a Bausch and Lomb Spectronic 20 spectrophotometer,

(4) cations (Ca^{++}, Mg^{++}, Na^{+}, and K^{+}), by A.A.S. as described earlier and

(5) silicon, by the molybdate blue method (Weaver et al., 1968).

Four soluble organic constituents were measured.

(1) Soluble carbon was determined by the Mebius method (Mebius, 1960), with reagent quantities adjusted for analyzing a 2 ml aliquot. The detection limit with these modifications was 7.5 mg l^{-1} of C.

(2) Soluble carbohydrates were measured by the anthrone method as modified by Oades (1967), with absorbance determined for this and the following two procedures on a Bausch and Lomb Spectronic 20 spectrophotometer.

(3) Total phenols were determined by the method of Swain and Hillis (1959), with the Folin-Denis reagent prepared according to the A.O.A.C. (1955). D-(+)-catechin was used as the standard.

(4) The carbazole method (Lynch et al., 1957) was employed for the

determination of uronic acids.

Freeze-dried samples of the August 4, 1979, collection of throughfall, stemflow, L-F, H, and Ah leachates were prepared and infrared spectra were produced by Spectral Services, Department of Chemistry, University of Alberta.

Litter traps 10 cm high, 55 cm square and with 2 mm mesh on the bottom, similar to those described by Cragg et al. (1978), were placed at ground level on 1.5 m^2 plastic sheets at 6 locations within an apparently undisturbed portion of the forest stand near the enclosure. Collections were made at intervals ranging from 1 to 6 weeks, depending on the season, from mid-May to late October, 1979. Litter materials were oven dried at 80^oC, sorted according to species and litter type, and stored for analysis.

Subsamples from the *Populus balsamifera*, *P. tremuloides*, and *Cornus stolonifera* leaf litter were taken on a weighted basis, allowing for the proportion of the year's total litterfall collected in each sampling period. These subsamples were combined for each species, ground in a Wiley mill and digested for elemental analysis by the method of Parkinson and Allen (1975). Ca^{++}, Mg^{++}, and K^+ contents of these digests were determined by A.A.S.

Litter bags (20 x 20 cm) were prepared in October, 1978, using freshly fallen leaves of the three species mentioned above. A mesh size of 3 mm was used in order to permit entry and feeding by soil mesofauna, as suggested by Gosz et al. (1973). Ten sets, consisting of one bag of each of the three species (5 g per bag) were placed at the base of the 1978 litter layer at the beginning of November. From the beginning of May, 1979, one set of bags was recovered each month until November of that year; a final collection was made in mid-March, 1980. The initial litter materials and

the monthly bag samples were dried, weighed, ground, and digested in the same fashion as the litter trap samples.

For comparison of the types and numbers of micro-organisms inhabiting the litter bag contents, one set was removed during mid-October, 1979. Subsamples were taken and dilution suspensions were prepared and plated out on plate count agar in Petri dishes which were incubated for 5 days at room temperature.

For comparison with the litter bag experiment, 2 g duplicate samples of the original leaf materials were ground to pass a 2 mm sieve and incubated according to the method described by I.A.E.A.-F.A.O.(1976). Evolved CO_2 was collected in NaOH and measured by titration with standardized HCl. The carbon content of the initial leaf materials prior to incubation was measured using the LECO induction furnace.

Intact aggregates from the Ckgj2 horizon were collected and oven dried. Features on the aggregate surfaces were examined with a Cambridge Stereoscan 150 scanning electron microscope and qualitative elemental analysis was performed with an associated Kevex 7000 Energy Dispersive X-ray Analysis Unit. No conductive coating was applied to these samples.

A final experiment was carried out at the site in order to observe the development of soil fabric characteristics in parent material samples subjected to conditions near the surface. Cylindrical cores (7.5 cm diameter x 8 cm long) were taken from the Ckgj2 horizon with a Uhland corer and inserted in the upper horizons at marked locations inside the fenced enclosure. Some of the cores were placed immediately below the L-F horizon, while the rest were completely enclosed within the Ah1 horizon. The cores were installed in September, 1977, and two were retrieved in October, 1979, one from each of the levels. The collected cores were oven dried and thin

sections were prepared in the same fashion as described earlier.

CHAPTER 4

HOLOCENE ENVIRONMENTAL HISTORY OF CENTRAL ALBERTA

4.1 Introduction

In keeping with the principal objective of this study, that is, determining if the pedon at the study site could have formed in its present environment, this chapter will review the evidence for Holocene environmental change in central Alberta. Beginning with the origin of the surficial deposits during the Wisconsinan glaciation, the discussion will consider evidence for Holocene climatic changes and their effects on vegetation zonation. The latter will be examined through a review of palynological studies, discussion of data on opal phytoliths in Alberta soils, and examination of historical records of pre-settlement vegetation.

4.2 Deglaciation and Surficial Geology

The present physiography of the Edmonton area contains elements attributable to both the preglacial landscape and its modification by glaciation and postglacial geomorphic processes (Westgate, 1969). The underlying bedrock throughout most of the Edmonton area belongs to the Upper Cretaceous Edmonton Formation, consisting of bentonitic shales and sandstones. For the township of interest to this study (25 - 51 - W4), Carlson (1964) indicated that the bedrock topography consists of a gentle northward slope toward the North Saskatchewan River. The landform that covers much of the Edmonton area, including the Ellerslie Research Station, was referred to as the Lake Edmonton Plain by Westgate (1969).

Lake Edmonton is the most common name for one of the proglacial lake phases that occurred as the Laurentide ice sheet retreated to the northeast at the end of the Wisconsinan glaciation. As discussed by St. Onge (1972), the deglaciation of north-central Alberta occurred rapidly, with ice-free conditions spreading from southwest of the Swan Hills c. 13,500 BP to Lofty Lake by c. 11,400 BP, a distance of about 300 km.

A complex series of short-lived proglacial lakes was continually changing as new outlets were exposed at progressively lower elevations. St. Onge (1972) identified two lake phases which occupied the Edmonton area, Lakes Leduc and St. Albert, corresponding to different levels controlled by the deepening of the Gwynne outlet. The southeastern extension of this outlet forms part of the present course of the Battle River. The duration of these phases controlled by the Gwynne outlet was quite brief; Bayrock and Hughes (1962) cited varve evidence for a period of about 40 years. Additional radiocarbon control for the time of deglaciation in the Edmonton area was given by Emerson (1977), who dated freshwater mollusc shells from supraglacial glaciolacustrine sediments in the Cooking Lake moraine, an area of ice stagnation. Ages ranged from 9050 $\pm$ 150 BP (I-4552) to 10,900 $\pm$ 190 BP (GSC - 2404).

The surficial deposits of the Lake Edmonton Plain vary greatly in both thickness (0-30 m) and texture (sand to clay) (Bayrock and Hughes, 1962). The proglacial nature of these deposits is indicated by the widespread occurrence of ice-rafted till and pebbles. Normally, the thinner deposits consist of varved silts and clays, with an upper clayey layer. In the vicinity of the study site, drill logs generally indicate this upward fining of the Lake Edmonton deposits, with thicknesses varying from approximately 20 m near the North Saskatchewan River, to about 5 m

immediately east of Blackmud Creek (McPherson and Kathol, 1972). In most cases, these logs showed the glaciolacustrine sediments to overlie till or occasionally bedrock.

From the evidence presented, it is likely that conditions were suitable for pedogenesis to begin on the Lake Edmonton Plain by about 10,000 BP.

4.3 Paleoclimates of the Holocene

The climatic conditions that would have influenced early Holocene and subsequent pedogenesis are incompletely understood. For North America as a whole, Bryson et al. (1970) surveyed radiocarbon dates related to episodes of climatic change and found a clustering corresponding to division points in the European Blytt-Sernander postglacial chronology. Their conclusion was that this agreement justified adoption of the scheme in North America. The middle part of this sequence, the Atlantic episode, was placed between 4680 and 8450 BP and essentially corresponds to the Altithermal, a period of warmer and drier conditions (Reeves, 1973).

Other paleoenvironmental studies both in Alberta and other parts of North America have produced a range of dates for the Altithermal. Such discrepancies are probably inevitable, given the variety and imprecision of the techniques employed, and the fact that climatic changes are not necessarily simultaneous or the same throughout a continent. As an example of this imprecision, Fritz and Krause (1973) found that oxygen isotope determinations on mollusc and ostracod shells indicated a period of accelerated evaporation from Lake Wabamun during the Holocene. No radiocarbon control was present, but since this episode occurred roughly in the middle of the sedimentary sequence, it was correlated with the Altithermal. For central Alberta in general, evidence for lowered water

levels or complete drying up of lakes suggest a drier and warmer climate during the approximate period 8700-6000 BP. The following 1000 years saw rapid change towards cooler and moister conditions, as inferred from pollen evidence, with limited additional change up to the present (Prof. C. Scheweger, University of Alberta, personal communication). A final example from southwestern Alberta illustrates another application of mollusc shells as a paleoenvironmental indicator and provides slightly different dates for the Altithermal. Harris and Pip (1973) used the habitat preferences of molluscs found in dated glaciolacustrine and fluvial deposits as evidence for warmer and drier conditions from 9000 to 7000 BP, followed by cooling.

4.4 Vegetation Change

Regardless of the imprecision of our understanding of Holocene climatic change in Alberta, it is clear that major fluctuations occurred and that these were reflected to some degree in vegetation patterns. From the standpoint of this study, it is important to know what the history of vegetation change has been in the Edmonton region so that its influence on soil genesis can be assessed.

Before discussing palynological evidence for vegetation fluctuations in central Alberta, it is important for comparative purposes to establish just where the present vegetation zone boundaries are located. Several authors have presented classifications and maps of vegetation types in the province (Bird, 1961; Moss, 1932, 1955; Rowe, 1972), but the most recent and systematic study was conducted by Zoltai (1975). Based on a detailed reconnaissance of the southern limits of four conifer species, Zoltai identified a Parkland Boreal Forest Transition zone which occupies

much of central Alberta, including the Edmonton area (Figure 1). He considered that this zone has a greater affinity with the Boreal Forest than the Aspen Parkland and could be considered to be a part of the former zone. Zoltai's Transition zone appears to correspond to Moss' (1932) Poplar Area, a zone dominated by Populus tremuloides and P. balsamifera forest, with the latter species increasing in abundance northwards and on moister sites. Moss (1932, 1955) also noted the tendency of white spruce (Picea glauca) to replace poplar in older unburned stands.

Palynological studies in the western interior of Canada have been reviewed and synthesized by Ritchie (1976). During deglaciation, the period 13000 to 9500 BP saw direct occupation of new surfaces by the early Boreal spruce forest. Simultaneously, grassland moved northwards and during the period 9500 to 6500 BP, a 50 to 100 km wide belt of birch-dominated deciduous forest existed between the Boreal forest and grassland zones. From 6500 to 2500 BP, the forest boundary extended southward, replacing grassland and occupying essentially its present position by 2800 BP.

In Ritchie's (1976) review paper, only two studies were cited from Alberta, both by Lichti-Federovich (1970, 1972). Taken from two sites approximately 100 km northeast of Edmonton, the Lofty Lake and Alpen Siding Lake cores present a similar sequence of pollen zones which generally follows the pattern outlined by Ritchie. One departure from the typical sequence is an early (c. 11,400 BP) and shortlived poplar maximum followed by a spruce-dominated assemblage until c. 9,200 BP. The succeeding zones, characterized by a decline in spruce pollen and an increase in birch, terminated at c. 7,500 BP. Following this is a period lasting until about 3,500 BP in which birch fluctuated in abundance, spruce in-

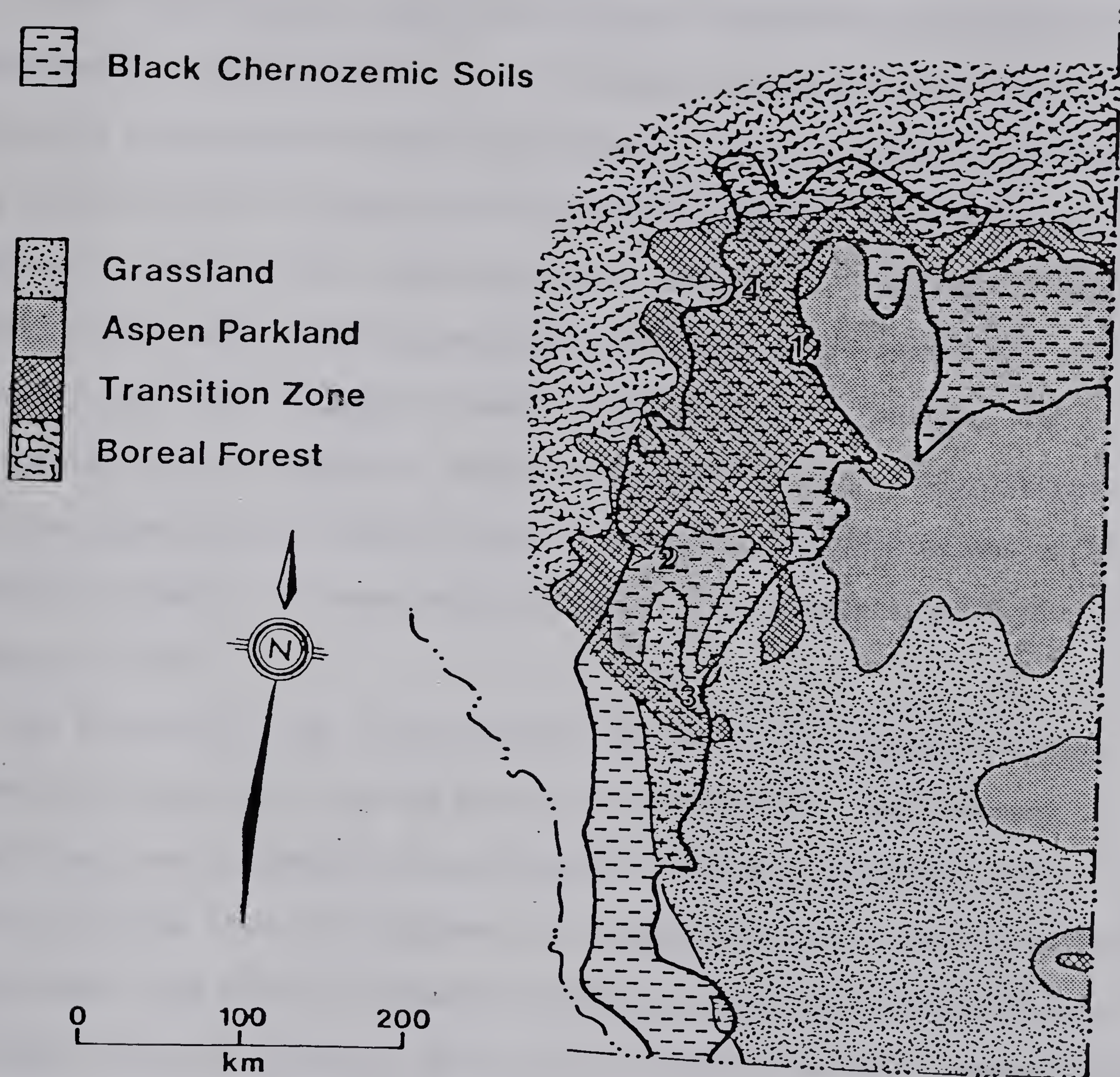

Figure 1. Vegetation zones of southern Alberta (after Zoltai, 1975). Phytolith sampling locations: 1 - Hay Lakes, 2 - Olds, 3 - Calgary, and 4 - Edmonton.

creased after 5200 BP, and grass pollen peaked before 5200 BP. The final zone suggested that little change in vegetation has occurred between 3500 BP and the present; arboreal pollen types dominated (spruce, birch, and alder).

Lichti-Federovich's (1970) paleocological interpretation supports the idea of an Altithermal interval, during which warming and decreased precipitation reached their maximum development c. 6000 to 5500 BP, followed by a deterioration in climate favouring Boreal species at the expense of grassland. However, the evidence does not suggest that grassland occupied the Lofty Lake area since the proportion of non-arboreal pollen was considerably less than is found in modern grassland pollen spectra. Two interpretations are suggested, involving dominantly birch-poplar forests with local grasslands on drier sites, or simply a northward extension of grassland so that its closer proximity increased the contribution of non-arboreal pollens.

The foregoing, then, casts doubt on the idea that a major northward extension of grassland occurred during the Altithermal. More recent unpublished work in central Alberta based on both pollen studies and evidence of lake level fluctuations, has suggested that Holocene vegetation fluctuations have been less dramatic in Alberta than in the other Prairie provinces (Prof. C. Schweger, University of Alberta, personal communication). Indications of falling water levels, brackish conditions, or even complete drying up of lakes point out a trend to drier and warmer conditions in the 8700 to 6000 BP period. However, pollen data from these studies do not suggest great changes in upland vegetation during that time, although some northward movement of the grassland zone is assumed.

Two recent pollen studies from the Cooking Lake moraine, approximately

30 km east of Edmonton, tend to confirm the suggestion that vegetation patterns have undergone little change during much of the late Holocene. A sediment core from Hastings Lake with a basal radiocarbon date of 4500 ± 190 BP showed an early peak abundance of grass, chenopod, and *Artemisia* pollen which declined in importance upward through the core (D. Emerson and Prof. C. Schweger, University of Alberta, personal communication). Cores from a pond and a bog with basal dates of 3970 ± 170 BP and 4180 ± 70 BP, respectively, in Elk Island National Park, showed similar pollen sequences (Vance, 1979). Both cores contained an initial zone in which grass and herb pollen equal or almost equal their abundance in contemporary grassland spectra in Manitoba. However, this zone was short-lived and tree and shrub pollen rapidly became dominant, although a major element of Gramineae persisted for about 1000 years. At about 2800 BP, *Betula* pollen increased sharply in abundance in both cores, parallelling a similar increase c. 3700 BP found by an unpublished pollen study carried out in Smallboys Lake, 90 km west of Elk Island Park. This change yielded a spectrum similar to that of the present vegetation and its regional occurrence suggests that a climatic factor, such as an increase in precipitation, was responsible. Vance's conclusion was, therefore, that vegetation similar to that of the present was established in the Elk Island Park area by c. 2800 BP.

While existing palynological data have given a tentative regional picture of Holocene vegetation change in central Alberta, site-specific information is desirable for this study. Numerous investigators in other parts of North America have examined opal phytoliths in the soil itself to provide that type of evidence (e.g. Norgren, 1973; Wilding and Drees, 1968; Witty and Knox, 1964). These structures are amorphous silica infillings

of plant cells, usually silt-sized, that persist in the soil after decomposition of the surrounding tissue. Although recent work has examined the forms of opal present in tree species (Geis, 1973; Klein and Geis, 1978), most emphasis has been given to opal phytoliths produced by grasses, simply because these plants have a higher silica content, between 10 and 20 times that found in legumes and other dicots (Jones and Handreck, 1967). As a result, where vegetation zone boundaries have remained stable, much higher contents of plant opal occur in grassland soils than in forest soils (Beavers and Stephen, 1958; Witty and Knox, 1964).

Of course, factors other than simple duration of grassland vegetation can affect the content of plant opal in a particular soil. Apart from obvious factors such as erosion and eolian influx, Jones and Beavers (1964) found that plant productivity, as related to drainage conditions, affected opal contents of surface soils. Parent material characteristics, notably the content of readily weathered silicate minerals, also govern the plant opal content in soils (e.g. Norgren, 1973).

For these reasons, a regional base of comparative data is needed before phytolith studies can be effectively used in reconstructing the vegetation history of a site. Unfortunately, little such background information exists for Alberta or the other Prairie provinces. Existing studies in this region have used phytoliths as a tool only in special situations, to identify possible paleosols (Dormaar and Lutwick, 1969) or to indicate the origin of slope deposits (Lutwick and Johnston, 1969).

To provide comparative data to that obtained for phytoliths from the Ellerslie site, three Ah horizon samples were obtained from Orthic Black Chernozems in other parts of the province: (1) the Hay Lakes area (Cooking Lake moraine), (2) the Olds area, and (3) a site approximately

15 km east of Calgary.*

The most striking pattern in the results (Table 1) is the southward increase in total opal phytolith content, with the Calgary sample showing about 10 times the amount present in the Ellerslie and Hay Lakes samples; the Olds sample lies in the middle of the range (Plates 1 and 2). The differences according to morphology are less consistent and are obscured by the large errors resulting from the low opal contents of some samples. At all sites, at least half of the 5-20 μm fraction opals are of the Elongate type which is found in all grasses (Table 2). Chloridoid types are present in only the Calgary and Ellerslie 5-20 μm fractions, but because of the very small opal content at the latter site, the 14% figure is likely a fortuitous result of a very low count; note the large error term. Since Chloridoid types are characteristic of short grass species (Twiss et al., 1969), they would be expected to be more abundant at the southerly site. Festucoid and Panicoid types characterize humid region grasses and species of the tall grass prairie, respectively, and were found only in the two southerly sites (Twiss et al., 1969). Hookbases were found in the 20-50 μm fractions at all sites and are not specific to any group of grasses (Norgren, 1973).

Distinctive plant opal forms derived from deciduous trees tend to be concentrated in the fine silt (2-5 μm) fraction (Wilding and Drees, 1974). Scanning electron microscope examination of a < 2.30 sp gr separate of this fraction from the Ellerslie Ah1 horizon revealed a variety of forms typical of deciduous tree leaves. Such forms were first identified in

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* (1) is from site 4, Edmonton area, and (2) is from site 1, Olds area, as designated by Dudas (1968).

Table 1 OPAL PHYTOLITH CONTENT OF SELECTED BLACK CHERNOZEMIC Ah HORIZONS

Site #	Location	5-20 μm*	20-50 μm	Total*
1	Hay Lakes	0.78	0.05	0.83
2	Olds	2.80	0.40	3.20
3	Calgary	5.36	0.64	6.00
4	Ellerslie	0.32	0.16	0.48

* Expressed as percentage of total soil.

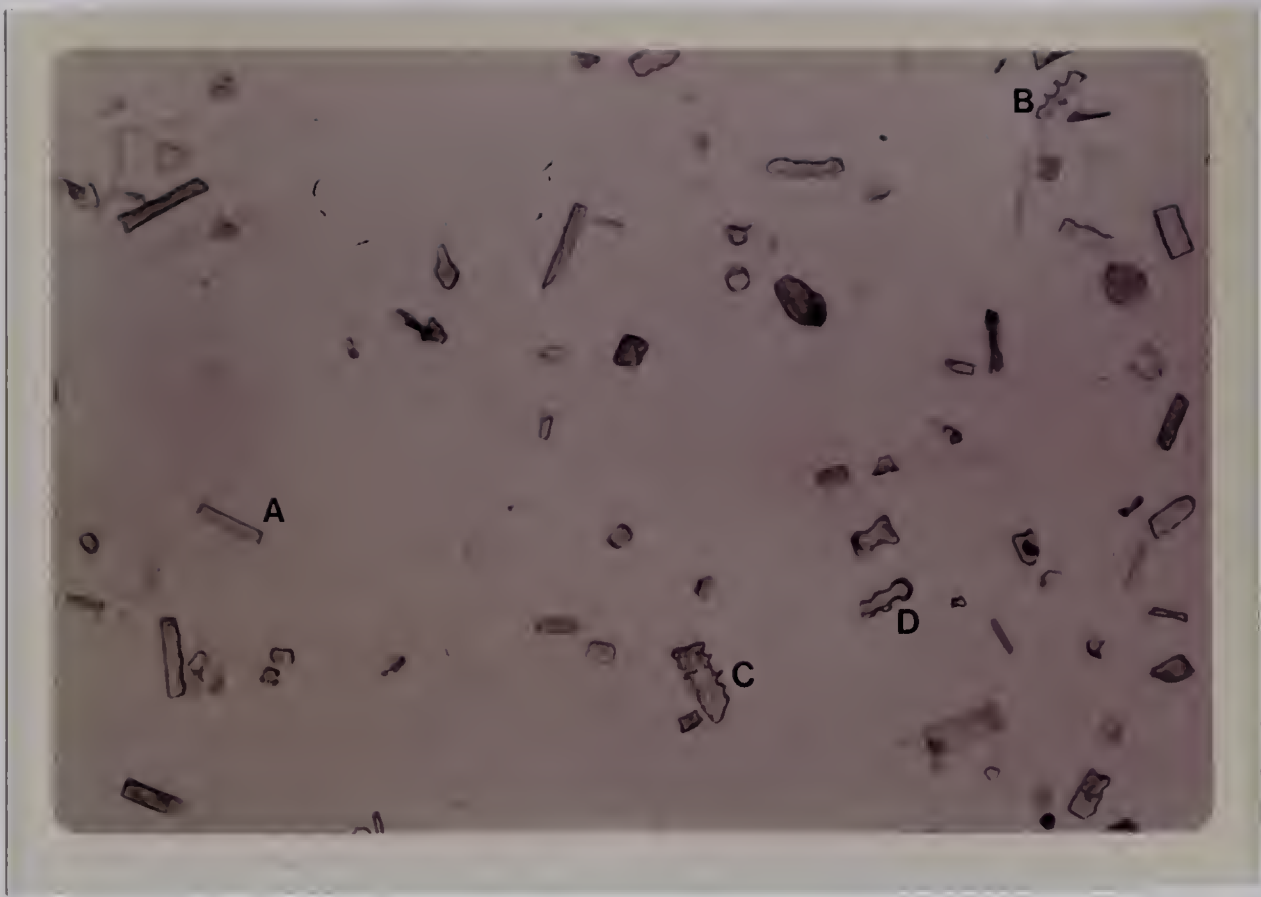


Plate 1. Calgary site, 5-20 μm silt (80 x). Phytolith types: A - smooth elongate, B - spiny elongate, C - rough elongate, D - panicoid.

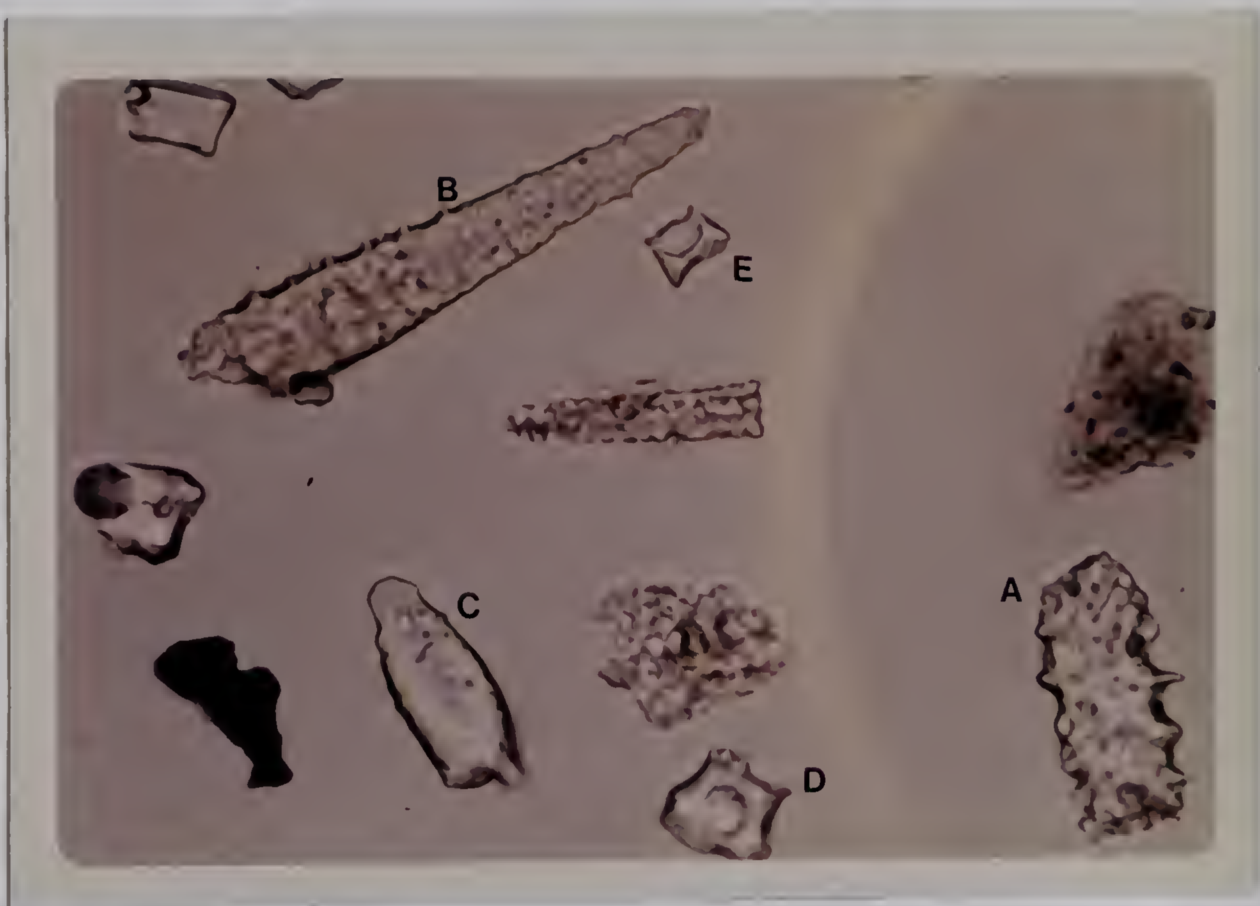


Plate 2. Calgary site, 5-20 μm silt (200 x). Phytolith types: A - spiny elongate, B - rough elongate, C - hookbase, D - chloridoid, E - chloridoid or panicoid (uncertain because of orientation).

Table 2

MORPHOLOGICAL ASSEMBLAGES OF OPAL PHYTOLITHS[1]

			Elongate									
Site No.	Location	Fraction	Smooth	Wavy	Rough	Spiny	Chloridoid	Festucoid	Panicoid	Hookbase	Other	n[2]
1	Hay Lakes	5-20 μm	21(33)	4(8)	29(19)	4(8)	-	-	-	13(14)	29(19)	24
		20-50 μm	-	-	-	-	-	-	-	100(0)	-	2
2	Olds	5-20 μm	36(11)	3(4)	31(11)	-	-	7(6)	1(2)	1(2)	21(10)	72
		20-50 μm	38(34)	-	25(31)	-	-	-	-	38(34)	-	8
3	Calgary	5-20 μm	15(6.3)	5(4)	26(7.7)	1(2)	5(4)	8(5)	5(4)	5(4)	31(8.1)	129
		20-50 μm	19(20)	6(12)	13(17)	6(12)	-	-	-	44(25)	13(17)	16
4	Ellerslie	5-20 μm	14(17)	57(37)	-	-	14(17)	-	-	-	14(17)	7
		20-50 μm	25(43)	-	-	-	-	-	-	75(43)	-	4

[1] Expressed as percentages of total opal counted in a given size fraction. () encloses error at 95.4% confidence level. Terminology after Twiss et al (1969) and Norgren (1973).

[2] n = total opal grains counted

several hardwood species from the eastern U.S. (Wilding and Drees, 1974), so the morphological types illustrated in Plates 3 and 4 are not associated with any particular species.

Any interpretation of vegetation history based on these data must necessarily be tentative in view of the small sample and lack of background information on the regional and between-species variations in opal content of Alberta grasses. However, given the observed dominance of grass-derived types, it does appear that total phytolith content could be related to the contribution of grasses to the vegetation cover during the history of these soils. This would suggest a much smaller grassland influence at the two Edmonton area sites than at the two more southerly sites. This relationship is even more significant when the site locations are compared to the contemporary vegetation zones delimited by Zoltai (1975) (Figure 1). Note that the Calgary area site is on the boundary between grassland and an extension of the Transition forest, while the Olds site lies between the Transition zone and the Aspen Parkland. This contrasts with the two Edmonton area sites which are both within the Transition forest zone. Such a correspondence between the apparent degree of grassland influence as indicated by phytoliths and the present vegetation zonation tends to suggest some stability in the latter.

A widely held view has been that the boundaries of the Boreal Forest and the grassland zone in the Prairie provinces have changed considerably since European settlement. Bird (1961) compared a vegetation map of Manitoba produced by Seton in 1905 with the present distribution of Aspen Parkland and concluded that there had been major southwestward expansion of the zone during this century. However, Zoltai (1975) re-examined the same area and disputed this conclusion:

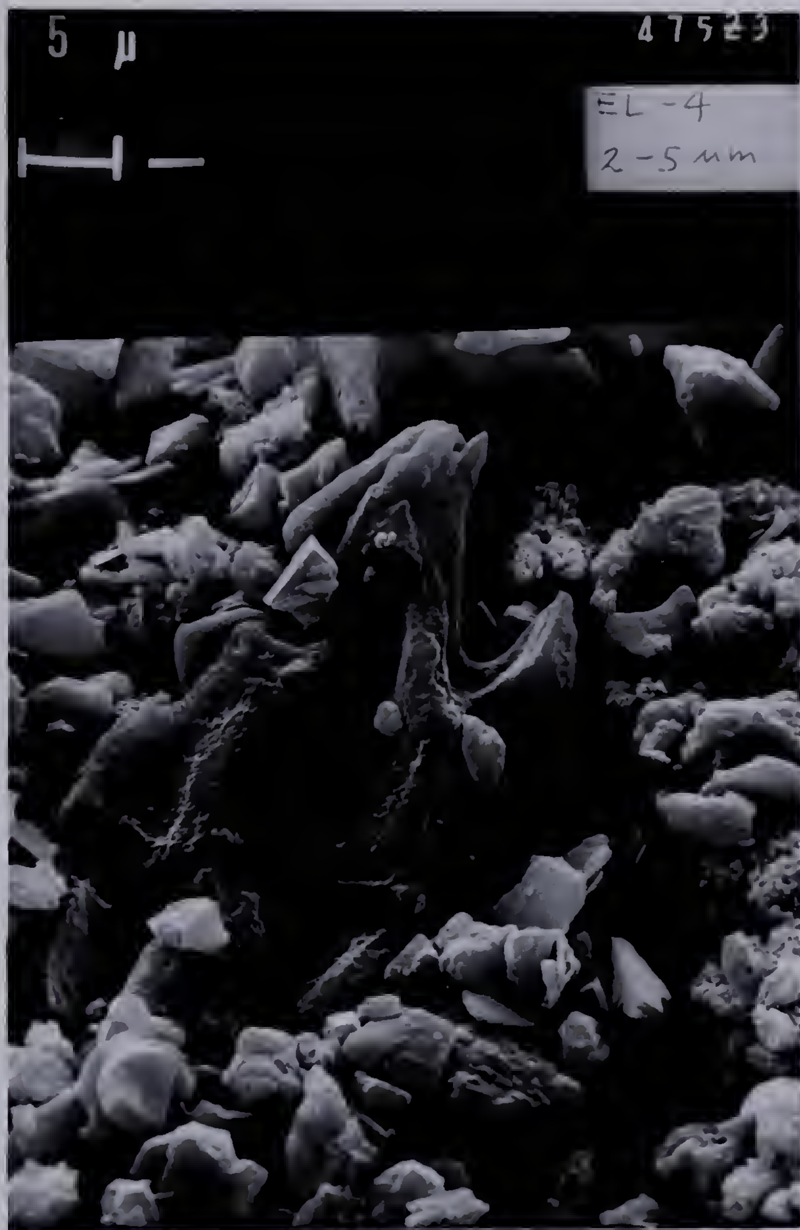


Plate 3. Scanning electron micrograph of deciduous tree-derived phytolith, Ellerslie site, Ah1 horizon. Structure is composed of opal encrustations on the walls of adjacent cells (probably epidermal).

Plate 4. Scanning electron micrograph of deciduous tree-derived phytolith, Ellerslie site, Ahl horizon. Note the layered structure.

> Many white and black spruce trees now growing in the area mapped as deciduous by Seton in 1905 were found to be over 100 years old and were alive at the time of his investigation. Similarly, many black spruce bogs occur within the "Deciduous Forest"; these bogs must have existed 70 years ago. It seems that early accounts must be examined critically before accepting them at face value. Vegetation changes are occurring continuously, but at a much slower rate than suggested by Seton's map.

While Zoltai's caution about early records is reasonable, it is still surprising that the original accounts by the Dominion Land Surveyors have not been more widely used in reconstructing pre-settlement vegetation. Elsewhere, such records have been a useful tool in studies of vegetation history (e.g. Rodgers and Anderson, 1979). For Alberta, such information is available both in the original surveyors' field notebooks and as township descriptions compiled from those notes (Dept. of the Interior, 1886). The only published application in Alberta of these data is by Johnston and Smoliak (1972), who compared an 1883 surveyor's notes with the present vegetation of a Parkland area north of the Porcupine Hills. Although a substantial recent expansion of aspen cover was indicated, it would be unwise to extrapolate this finding to the Edmonton area in view of the different vegetation zone involved.

From the 1886 compilation of township descriptions, it seems clear that most of the Lake Edmonton Plain was wooded, dominantly with poplar and lesser amounts of spruce. The other vegetation type widely reported was a scrubby willow-poplar mixture. The only major open areas described were usually marshy. For Tp. 51, R. 25, W4, which includes the Ellerslie Research Station in section 24, the 1883 surveyor gave the following description: "Well timbered with poplar, except a narrow strip about half a mile wide, north and southwards, which is covered with scrub and willow." The eastern edge of this township which forms the boundary of the Station,

was surveyed in 1883 and was described as "prairie, with a first class soil, covered with dense poplar and willow, intermingled with spruce" (Figure 2). The interior of that township was "Timbered with poplar and a few scattered spruce, among which there are openings covered with scrub and high willow and poplar scrub. There is no open prairie and the soil is generally good." Apparently the term "prairie" did not apply to a vegetation type, but merely denoted flat land.

Earlier travellers in the 19th Century described the Fort Edmonton area in similar terms. Approaching from the east in 1859, the Earl of Southesk "camped for the night on a knoll a few hours from Edmonton, from which there was a beautiful view over a circle of wooded plain, perfectly level except where the steep north bank of the river was discernible" (Southesk, 1969). The year before, James Hector of the Palliser expedition had observed that

> Edmonton must be considered as being in the wooded country, but in the immediate neighbourhood of the fort there is not much valuable timber....Once back from the river banks, which are everywhere high and precipitous, the country is rather flat, and covered with thickets of willow and poplar, and with a much larger proportion of swampy ground than I have seen elsewhere in the Saskatchewan. (Spry, 1968)

As described by Zoltai (1975) and the Dominion Land Surveyors, the native vegetation of the Lake Edmonton Plain corresponds well to the remnant forest stand occupying the study site (Table 3). The stand of _Populus balsamifera_ does not have a complete canopy, but apparently because of its age (80-85 years) it is opening up as older trees fall over (Plate 5). There is little regeneration of the species, with most of the younger trees being _P. tremuloides_, particularly at the edge of the stand. The stand is described as being at an intermediate successional stage, with the next dominant that would establish itself being _Picea glauca_, pro-

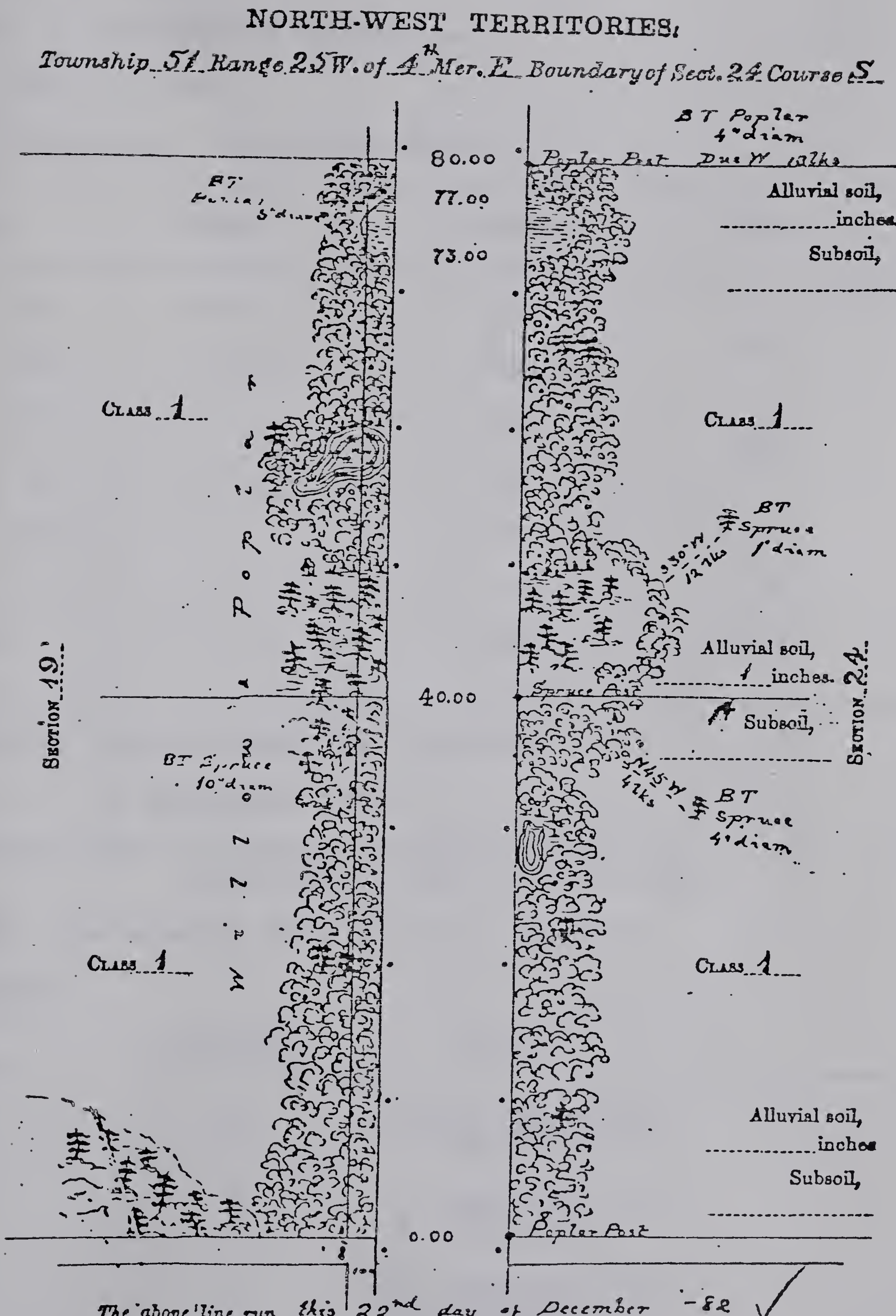


Figure 2. Dominion Land Surveyor's notebook entry, 1882.
Ellerslie Research Station is on right side of survey line.

Table 3 VEGETATION DESCRIPTION, ELLERSLIE STUDY SITE

Plot Size: 20 x 20 m

Location: 5 m southeast of site enclosure

Date: August 17, 1978

General Physiognomy: Balsam Poplar Forest

Stratum		Height	Coverage	Total
Tree:	A1	16-20 m	15%	25%
	A2	5-16 m	10%	
Shrub:	B1	2-5 m	12%	100%
	B2	< 2 m	90%	
Herb:	Ch		25%	25%
Mosses			< 1%	< 1%
Lichens			< 1%	< 1%

Regeneration: Populus tremuloides - moderate

P. balsamifera - weak

Successional Stage: Intermediate, probably succeeding to Picea glauca if seed source available.

Stand Age: Approximately 80-85 years (P. balsamifera)

Species List*

Layer	% Coverage	Species
A1	14	Populus balsamifera L.
	1	P. tremuloides Michx.
A2	6	P. tremuloides Michx.
	4	P. balsamifera L.
B1	7	P. tremuloides Michx.
	3	P. balsamifera L.
	1	Prunus virginiana L.
	+	Salix sp.
	+	Amelanchier alnifolia Nutt.

(Table 3 continued)

Layer	% Coverage	Species
B2	65	Cornus stolonifera Michx.
	15	Rosa acicularis Lindl.
	5	Symphoricarpos albus (L.) Blake
	5	Rubus strigosus Michx.
	4	Viburnum edule (Michx.) Raf.
	3	Lonicera involucrata (Richards.) Banks
	3	Ribes triste Pall.
	2	Lonicera dioica L. var. glaucescens (Rydb.) Butters
	1	A. alnifolia Nutt.
	1	P. balsamifera L.
Ch	6	Rubus pubescens Raf.
	2	Mitella nuda L.
	2	Mertensia paniculata (Ait.) G. Don
	2	Cornus canadensis L.
	1	Heracleum lanatum Michx.
	1	Galium boreale L.
	1	G. triflorum Michx.
	1	Aster conspicuus Lindl.
	1	Lathyrus ochroleucus Hook.
	1	Epilobium angustifolium L.
	+	Urtica gracilis Ait.
	+	Cichorium intybus L.
	+	Geranium sp.
	+	Fragaria virginiana Duchesne
	+	Viola renifolia A. Gray
	+	Calamogrostis canadensis (Michx.) Beauv.
	+	Vicia americana (Muhl.)
	+	Aster ciliolatus Lindl.
	+	Smilacina stellata (L.) Desf.
	+	Agropyron repens (L.) Beauv.
	+	Thalictrum venulosum Trel.
	+	Solidago multiradiata Ait..
	+	Geum macrophyllum Willd.
	+	Taraxacum officinale Weber
	+	Pyrola asarifolia Michx.

* Nomenclature after Moss (1959). + indicates < 1%.

Plate 5. Balsam Poplar forest adjacent to study site enclosure. Shrub layer is dominantly <u>Cornus stolonifera</u>.

vided that a seed source is available (Dr. I. Corns, Canadian Forestry Service, personal communication). However, no spruce seedlings were observed, which is not surprising, since the nearest mature individuals are 800 m away across cultivated fields.

A major feature of the site vegetation is the prominent shrub stratum (< 2 m) in the understorey. Although dominated by *Cornus stolonifera*, there is great diversity, with 10 species in this layer. The herb layer is also diverse, with 25 species represented, and has a coverage of about 25%.

This discussion has emphasized the interpretation of vegetation changes and patterns in relation to climatic influences, yet at the more local level, edaphic factors are important. For example, the Dominion Land Survey records indicated that the only extensive open, non-forested vegetation in the Edmonton area occurred as poorly-drained meadows. Moss (1932) noted that such sites are commonly occupied by an association of Salix spp. and *Calamogrostis canadensis* and that with drier conditions, poplars are able to invade. As noted earlier, *P. balsamifera* is generally found on moister sites, so its dominance at the study site is consistent with the emphasis in the historical accounts on the large amount of poorly-drained land and the abundance of willow in the Edmonton area. Solonetzic soils are found within dominantly Chernozemic landscapes in central Alberta, yet as noted by Moss (1955), their native vegetation has received little study. Judging from personal observations of the areas of Solonetzic soils east of the Cooking Lake moraine, growth of poplar is poor on such sites. Any past influence of salinity, resulting from groundwater discharge at the study site, may have promoted a more open vegetation cover. For example, Moss (1955) did not consider the possibility

that the existence of parkland vegetation in the Peace River region may be related to the widespread occurrence of Solodic soils.

4.5 Summary and Conclusions

This review of Holocene environmental history in central Alberta leads to three principal conclusions relevant to this study.

(1) Pedogenesis probably began at the study site c. 11000-10000 BP following deglaciation and a brief inundation by proglacial Lake Edmonton.

(2) Paleoclimatic evidence indicates warming during the early Holocene, culminating in the Altithermal episode at around 6000 BP, followed by progressive cooling and moister conditions up to the present time.

(3) These climatic fluctuations have been accompanied by both migration of vegetation zones and changes in their composition. However, it is still unclear whether there was a prolonged northward invasion of grassland into central Alberta during the Altithermal. Opal phytolith evidence suggests that the duration of any such invasion was much briefer in the Edmonton area than in the Aspen Parkland to the south. In any event, pollen studies demonstrate that there has been little change in vegetation composition during the past 3000 years. Historical evidence confirms that the Lake Edmonton Plain was forested at the time of European settlement and indicates that the forest composition fitted the concept of Zoltai's (1975) Transitional zone between the Boreal Forest and the Aspen Parkland.

CHAPTER 5

PEDON CHARACTERISTICS AND GENETIC PROCESSES

5.1 Introduction

This chapter will present and discuss the main body of data from this study. Chapter Two has outlined the central genetic concepts of Chernozemic soils and, as was pointed out, these ideas have developed largely from inferences made from their properties and environmental relationships. In keeping with this traditional approach, the first part of this chapter will treat the study site pedon in terms of its properties (morphological, physical, chemical, and mineralogical) and the genetic inferences that can be drawn from them.

The second part of the chapter will take a complementary approach, presenting the results of process studies carried out at the site: monitoring of the physical environment, nutrient cycling, litter decomposition, solute dynamics in water entering and moving through the soil system, and fabric changes occurring near the soil surface in parent material samples.

5.2 Pedon Characteristics

5.2.1 Macro- and Micromorphology

Based on morphological and analytical properties, the pedon under investigation is classified as a Gleyed Eluviated Black Chernozem (Canada Soil Survey Committee, 1978). The common horizon sequence indicated for this subgroup (Ah, Aej, Btjgj, Ckgj) essentially fits, although there did not appear to be sufficient textural and structural development to warrant

designation of a Btjgj horizon (Table 4).

Unlike the common horizon sequence, however, this pedon is distinguished by a prominent 18 cm thick set of organic horizons (L-F, F-H, and H) (Plate 6). According to the traditional nomenclature for humus forms, this grouping characterizes the moder type (Bernier, 1975). A high content of fecal pellets was evident in these horizons, both in the field and in thin section; in the latter, these pellets comprised the humigranic units (Table 5, Plate 7). Two dominant size classes of fecal pellets were noted in the thin section of the L-F horizon: 0.25-0.75 mm in diameter, which were randomly distributed, and approximately 50 μm in diameter, which were concentrated within or beside leaf residues. These fecal materials became most abundant toward the base of the H horizon, while there was a corresponding decrease in the content of recognizable plant residues (foliohists and brancohists). In addition, more mineral material was present in the H horizon, producing a mull-humigranoidic fabric.

The Chernozemic A horizon of the pedon constitutes a second humus form, mull, characterized by intimate mixing and aggregation of humic and mineral materials (Bernier, 1975). Based on colour (indicative of organic matter content), structure (granular, becoming finer with depth), and root abundance (decreasing with depth), three Ah horizons were distinguished. Fabric types are dominantly mullgranoidic with areas of mullgranic fabric associated with root or faunal channels (Plate 8). The structural units range between 0.1 and 1.0 mm in diameter and vary in their degree of coalescence. The Ah horizons were underlain by an Ahej horizon distinguished by colours indicative of lower organic matter content, fewer roots, and a transition from granular to subangular blocky

Table 4 PEDON DESCRIPTION

Horizon	Depth (cm)	Description
L-F	18-13	Dark brown (7.5 YR 3/2 m) partially decomposed Populus balsamifera leaves, with few partially decomposed roots; abundant medium fine and very fine random and horizontal roots; 2 to 6 cm thick; abrupt, smooth boundary; pH 7.2.
F-H	13-5	Dark reddish brown (5 YR 3/3 m) humified and partially humified organic matter; loose and fluffy; abundant coarse and medium horizontal, fine and very fine random roots; clear, smooth boundary; 5 to 9 cm thick; pH 6.9.
H	5-0	Dark red brown (5 YR 2/2 m) humified organic matter; loose; abundant coarse and medium horizontal, fine and very fine random roots; abrupt, smooth boundary; 4 to 6 cm thick; pH 6.6.
Ah1	0-10	Black (10 YR 2/1 m) silty clay; strong granular; slightly firm; common fine and very fine and few medium and coarse oblique inped and exped roots; gradual, wavy boundary; 8-11 cm thick; pH 6.1.
Ah2	10-20	Very dark gray (10 YR 3/1 m) silty clay; strong medium and fine granular; slightly firm; common fine and very fine oblique inped and exped roots; clear wavy boundary; 7 to 9 cm thick; pH 6.4.
Ah3	20-29	Dark gray (10 YR 4/1 m) to very dark gray (10 YR 3/1 m) silty clay; strong fine granular; slightly firm; few very fine random inped and exped roots; clear wavy boundary; 7 to 9 cm thick; pH 6.8.
Ahej	29-35	Very dark grayish brown (2.5 Y 3/2 m) silty clay; strong granular to fine subangular blocky; firm; very few fine and very fine random inped and exped roots; abrupt, wavy boundary; 4 to 7 cm thick; pH 6.7.
Aegj	35-42	Grayish brown (2.5 Y 5/2 m) silty clay; many, fine, prominent (2.5 Y 5/6) mottles; very weak fine platy to weak fine granular; firm; very few fine and medium random inped and exped roots; abrupt, broken boundary; 0 to 8 cm thick; pH 6.9.

Table 4 continued

Horizon	Depth (cm)	Description
Bmgj	42-57	Dark grayish brown (2.5 Y 4/2 m) silty clay; common fine, distinct (2.5 Y 4/4) mottles; moderate fine subangular blocky; very firm; very few fine and medium random inped and exped roots; gradual wavy boundary; 15 to 20 cm thick; pH 7.3.
Bmkgj	57-67	Dark grayish brown (2.5 Y 4/2 m) clay; common, fine, distinct (2.5 Y 4/4) mottles; weak subangular blocky; very firm; very few large oblique inped and exped roots; weakly effervescent; gradual, wavy boundary; 8 to 12 cm thick; pH 7.3.
BCkgj	67-103	Dark grayish brown (2.5 Y 4/2 m) heavy clay; many, fine, distinct (2.5 Y 4/3-4/4) mottles; massive to weak fine subangular blocky; very firm; very few, medium, fine, and very fine random and oblique roots; moderately effervescent; gradual, wavy boundary; 25 to 30 cm thick; pH 7.8.
Ckgj1	103-128	Dark grayish brown (2.5 Y 4/2 m) heavy clay; many, fine, prominent (2.5 Y 5/4-5/6) mottles; massive to weak subangular blocky; very firm; very few, medium, fine, and very fine random and oblique roots; moderately effervescent; gradual wavy boundary; 25 to 30 cm thick; pH 7.8.
Ckgj2	128 +	Dark grayish brown (2.5 Y 4/2 m) heavy clay; many, fine, prominent (2.5 Y 5/4-5/6) mottles; massive, fragmental; moderately effervescent; pH 7.4.

Plate 6. Gleyed Eluviated Black Chernozemic profile, Ellerslie study site.

Table 5. MICROMORPHOLOGICAL DESCRIPTION*

Slide No.	Horizon	Zone	Depth (cm)	Fabric	Compositional Remarks
1	L-F	I	13-10	Granic	Mixed complex fabric; humi-phytogranic; radicohists 0.5-3.0 mm diameter; two size groups of granic units: 0.25-0.75 mm and ≈ 50 μm, with the smaller units within or beside phytogranic units (foliohists); abundant fungal hyphae; rare sclerotia; undulic plasma fabric.
	F-H	II	10-9	Granic	Mixed complex fabric; phyto-humigranic; dominance of 50 μm humigranic units; phytogranic units (foliohists) thinner and smaller than in I; undulic plasma fabric.
	F-H	III	9-8½	Granic	Like I, but with lower content of humigranic units.
	F-H	IV	8½-7	Granic// granoidic	Mixed and separated complex fabrics; phyto-humigranic, coalescing to granoidic at base of zone; phytogranic units decreasing in abundance to base of zone; radicohists and brancohists 0.5-2.0 mm diameter; undulic plasma fabric.
2	F-H	I	7-2	Granoidic	Humigranoidic (50-100 μm diameter), some humigranic in voids; few phytogranic units (foliohists) radicohists and brancohists 0.5-2.0 mm diameter; undulic plasma fabric.
	H	II	2-½	Banded granoidic-fragmoidic	Mixed complex fabric; mull-humigranoidic-fragmoidic, separated by subparallel joint planes; more mineral grains than in above zones; argillasepic plasma fabric.

*Terminology after Bal (1973), Brewer (1964), and Brewer and Pawluk (1975).

Table 5 continued

Slide No.	Horizon	Zone	Depth (cm)	Fabric	Compositional Remarks
3	H	I	½-0	Banded granoidic-fragmoidic	Mixed complex fabric; mull-humigranoidic-fragmoidic, separated by subparallel joint planes; more mineral grains than in above zones; argillasepic plasma fabric.
	Ah	II	0-3½	Granoidic-porphyric	Mixed complex fabric; mull-granoidic (0.5-1.0 mm diameter) - vughy porphyric, zones of granic granoidic; skelmosepic plasma fabric, transitional boundary.
	Ah	III	3½-6½	Fragmoidic granoidic/granic	Separated complex intergrade fabric; metafragmoidic mullgranoidic/mullgranic (units 100-1000 μm); more open structure than in zone above; skelmosepic plasma fabric.
4	Ah		6½-13	Granoidic	Mullgranoidic fabric (units 100-1000 μm diameter) fabric; some planar voids, but mostly vughs, little accommodation; skelmosepic plasma fabric.
5	Ah	I	13-19½	Granoidic	Mullgranoidic fabric (units 100-1000 μm diameter); skelmosepic plasma fabric.
	Ah	II	14½-18½	Granic-granoidic	Mixed complex fabric; mullgranic-mullgranoidic; located in vertical channel; skelmosepic plasma fabric.
6	Ah		19½-26½	Porphyric granoidic//granoidic	Separated complex intergrade fabric; dominantly mullgranoidic with areas of vughy porphyric mullgranoidic; skelmosepic plasma fabric.

Table 5 continued

Slide No.	Horizon	Zone	Depth (cm)	Fabric	Compositional Remarks
7	Ahej	I	26½-30½	Granoidic-porphyric	Mixed complex fabric; mull-granoidic-vughy porphyric, random scattering of circular groups (0.5-1.0 mm diameter) of coalesced mull-granoidic units, possibly pedotubules; faint reddish-brown Fe nodules; skelmosepic plasma fabric; transitional boundary.
		II	30½-33	Porphyric	Vughy porphyric; otherwise like zone I.
8	$\overline{Aegj}$		33-40	Granoidic porphyric/ porphyric	Separated complex intergrade fabric; vughy porphyric, zones of granoidic porphyric; like II of slide 7, except that more of voids are skew planes; vertical root channels (< 1 mm wide) with melanotic radicohists; distinct 100-250 μm Fe nodules; masepic to skelmosepic with zones of omnisepic plasma fabrics.
9	Bmgj		40-47	Granoidic-fragmoidic porphyric/ porphyric	Separated and mixed complex intergrade fabric; granoidic-fragmoidic vughy porphyric/ vughy porphyric; vertical root channels (< 1 mm wide), some with melanotic radicohists; distinct 100-250 μm Fe nodules; masepic to mosepic, some skelsepic plasma fabric.
10	Bmgj	I	47-50	Granoidic-fragmoidic porphyric/ porphyric	Like slide 9, but with argillasepic, some masepic plasma fabric; transitional boundary to:
		II	50-54	Fragmoidic porphyric// porphyric	Separated complex intergrade fabric; closely packed metafragmoidic units grading to vughy porphyric; otherwise like I.

Table 5 continued

Slide No.	Horizon	Zone	Depth (cm)	Fabric	Compositional Remarks
11	Bmgj	I	54-56	Fragmoidic porphyric	Intergrade fabric; fragmoidic vughy porphyric; otherwise like slide 9; transitional boundary.
	Bmkgj	II	56-60½	Granoidic fragmoidic// porphyric	Separated complex intergrade fabric; granoidic fragmoidic// vughy porphyric; some granic-granoidic in pedotubules; faint to distinct Fe nodules (100-250 μm diameter); argillasepic, some masepic plasma fabric.
12	Bmkgj		60½-67½	Granoidic fragmoidic// porphyric	Ditto, with well developed aggrotubule containing melanons and mullgranoidic fabric.
13	BCkgj		67½-74	Fragmoidic porphyric// porphyric	Separated complex intergrade fabric; fragmoidic vughy porphyric// vughy porphyric; otherwise like slide 11, zone II
14	BCkgj		74-80½	Porphyric// fragmoidic porphyric	Separated complex intergrade fabric; vughy porphyric// fragmoidic vughy porphyric; vertical root channels, some with melanotic radicohists (0.5-1.0 mm wide); Fe nodules (100 μm diameter) both within and at margins of units; masepic, unistrial argillasepic plasma fabric.
15	BCkgj		80½-87	Ditto	Like slide 14, but with black (Mn) nodules (100-200 μm diameter).
16	BCkgj		87-93½	Fragmoidic porphyric// porphyric	Separated complex intergrade fabric; fragmoidic vughy porphyric// porphyric; small zone (2 x 0.5-1.0 cm) containing granic-granoidic fabric; otherwise similar to 15; masepic plasma fabric.

Table 5 continued

Slide No.	Horizon	Zone	Depth (cm)	Fabric	Compositional Remarks
17	BCkgj		93½-100½	Fragmoidic porphyric// porphyric	Like 16, with some granoidic fragmoidic fabric along major joint planes; elongated zones of sand and silt between some units.
18	BCkgj		100½-107½	Ditto	Ditto
19	Ckgj1		107½-114½	Fragmoidic/ fragmoidic porphyric/ porphyric	Separated complex intergrade fabric; dominantly vughy porphyric with minor areas of intergrade (fragmoidic vughy porphyric) and fragmoidic fabrics intermixed; vughy porphyric zones tend to have higher sand content; unistrial argillasepic, omnisepic plasma fabric.
20	Ckgj1		114½-121½	Fragmoidic porphyric// porphyric	Separated complex intergrade fabric; fragmoidic vughy porphyric// vughy porphyric; mangans along joint planes; omnisepic, unistrial argillasepic plasma fabric.
21	Ckgj1		121½-128	Ditto	Ditto, with granic fabric in major channels.
22	Ckgj2		128-135	Ditto	Ditto, with rare void argillans.
23	Ckgj2		135-142	Porphyric-fragmoidic	Mixed complex fabric; vughy porphyric-fragmoidic; distinct void neomangans; omnisepic, unistrial argillasepic plasma fabric.
24	Ckgj2		142-149	Ditto	Ditto, with rare void argillans.
25	Ckgj2		149-156½	Ditto	Ditto.

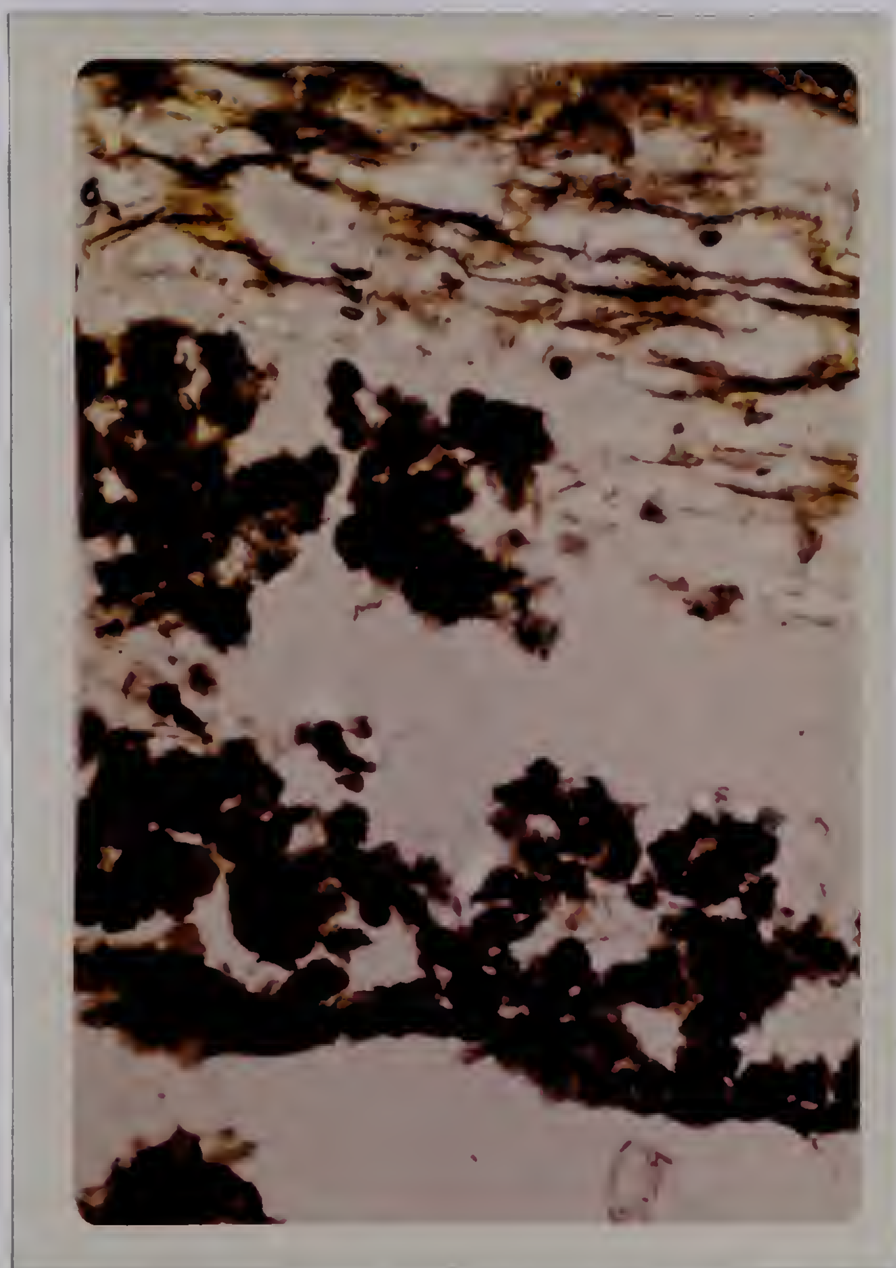


Plate 7. Thin section view of L-F horizon, with dominance of 50 μm diameter fecal pellets.

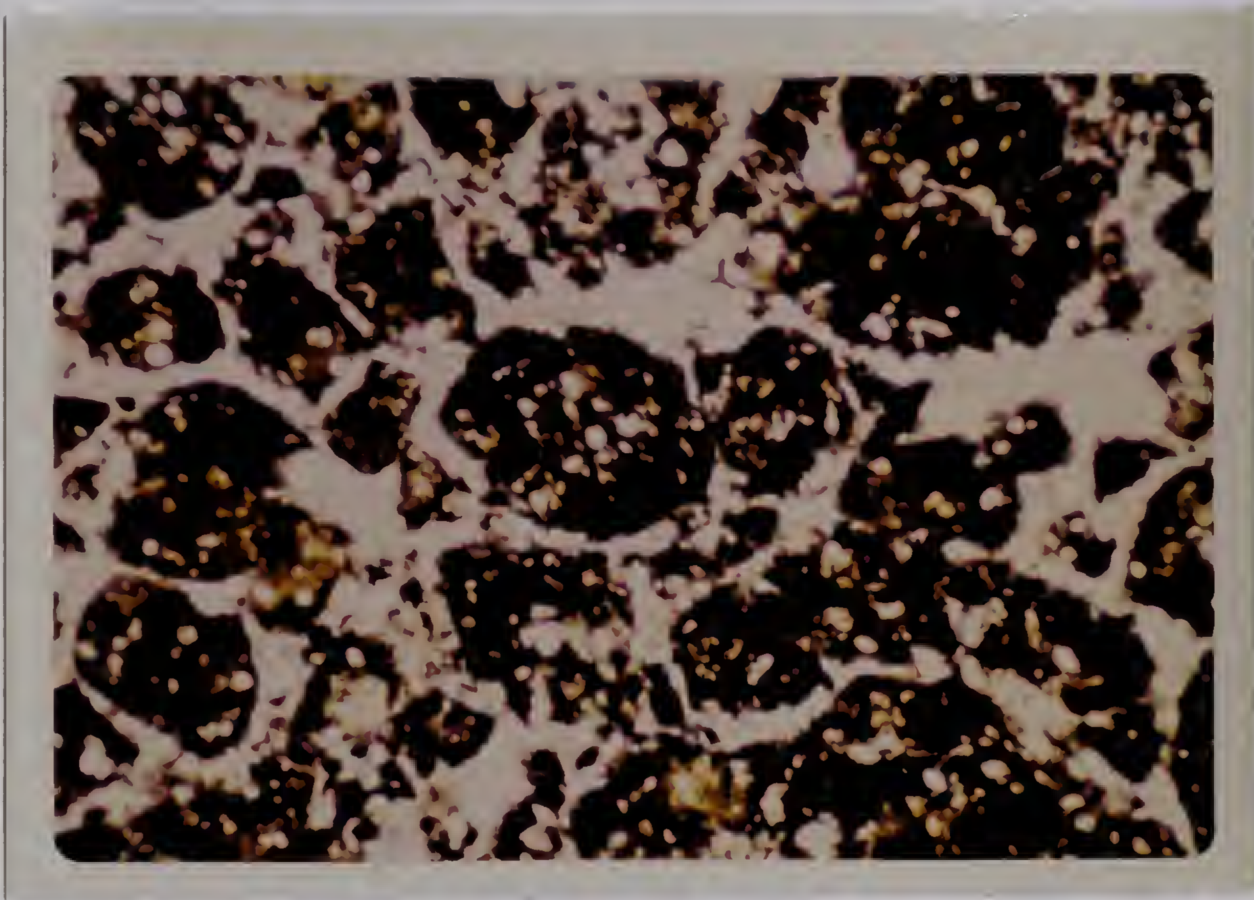


Plate 8. Thin section view of mullgranoidic fabric, Ah1 horizon. Units range from 0.2 to 1.0 mm diameter.

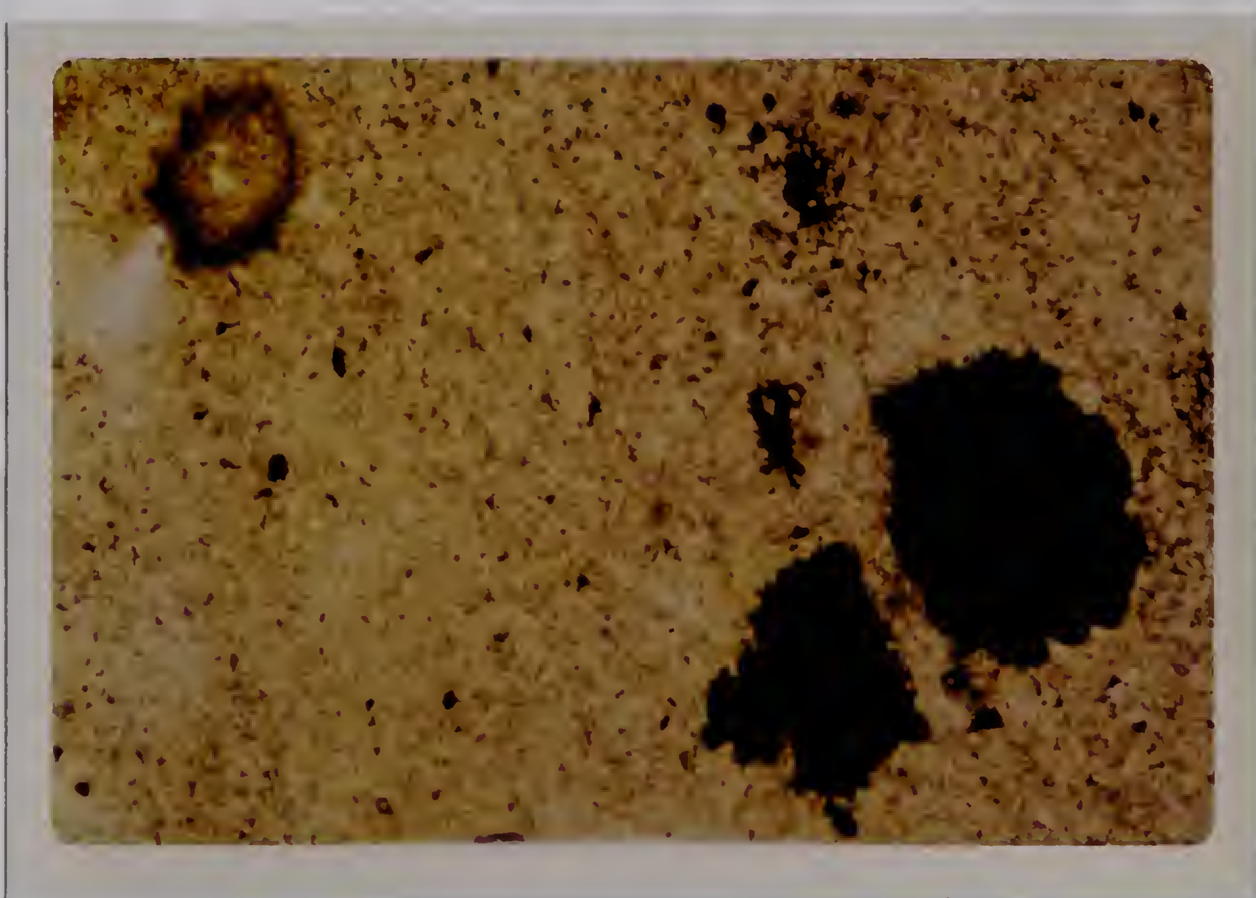


Plate 9. Thin section view of iron (reddish-brown) and manganese (black) nodules, BCkgj horizon. Lower manganese nodule is approximately 0.2 mm in diameter.

structure. Micromorphologically, this horizon had a denser fabric showing greater coalescence of units (mullgranoidic-vughy porphyric). Although mottles were not observed in the field, faint Fe nodules were visible in thin section. Plasma fabrics were skelmosepic in both the Ah and Ahej horizons.

The A$\overline{\text{egj}}$ horizon was discontinuous and weakly developed with no textural difference between it and the Bmgj horizon being evident in the field. The horizon was distinguished by its somewhat bleached appearance which contrasted with the darker overlying horizons of higher organic, matter content. Structural development was very weak and at times during the year, this horizon was very difficult to discern. Mottling was observed both in the field and in thin section.

The B horizon was divided into upper (Bmgj) and lower (Bmkgj) portions because of a weaker subangular blocky structure and the presence of weak effervescence in the latter. Mottles were less numerous than in the A$\overline{\text{egj}}$ horizon, although of similar size and colour. Fabrics differed from those in the Ah horizons, with a greater coalescence of structural units which were more angular, resulting in a dominantly vughy porphyric and fragmoidic character. Root channels and aggrotubules were evident. Plasma fabrics varied between argillasepic and insepic.

The transitional BCkgj and the underlying Ckgj horizons were characterized by a higher degree of effervescence, a disappearance of pedogenic structure, and a heavy clay texture. Mottles became more distinct with depth and thin sections showed them to be both adjacent to voids and within the structural units. Black stains were noted on joint plane surfaces, particularly in the Ckgj horizons, both in the field and in thin sections. An X-ray energy dispersive analysis of one such black zone on the surface

of a fragmental unit from the Ckgj2 horizon indicated a high content of Mn compared to the matrix; hence, these features were considered to be mangans (Figure 3, Plate 9). Fabrics varied between fragmoidic and vughy porphyric and local elongated zones contained higher contents of sand and coarse silt. The high clay content was expressed in plasma fabrics that were dominantly argillasepic and omnisepic.

5.2.2 Particle Size Characteristics

Although the textures in this pedon are all clay-rich (silty clay to heavy clay), textural class designations tend to obscure a considerable degree of variability (Table 6). Sand content ranged from 1.6% to 16.1%, while clay content varied between 41.5% and 63.7%. Textural variations showed no systematic relationship to genetic horizon type, with the highest fine clay/total clay ratios occurring in the supposedly eluvial horizons (Ahej and $\overline{\text{Aegj}}$). The only general trend was towards finer textures at the base of the pedon. While the coarse fragment content was virtually nil in this pedon, similar soils elsewhere at the Ellerslie Research Station occasionally contained discontinuous lenses of fine gravel and sand. Such a layer was encountered at 4.5 m during installation of the piezometers at the site.

5.2.3 Organic Carbon and Total Nitrogen

Both of these components generally decreased in abundance with depth (Table 6). The maximum values occurred in the organic horizons, with the highest organic carbon content in the L-F horizon and the highest total N in the H horizon. In each succeeding horizon, organic carbon content decreased by about half between the Ahl and Aegj horizons. C/N ratios

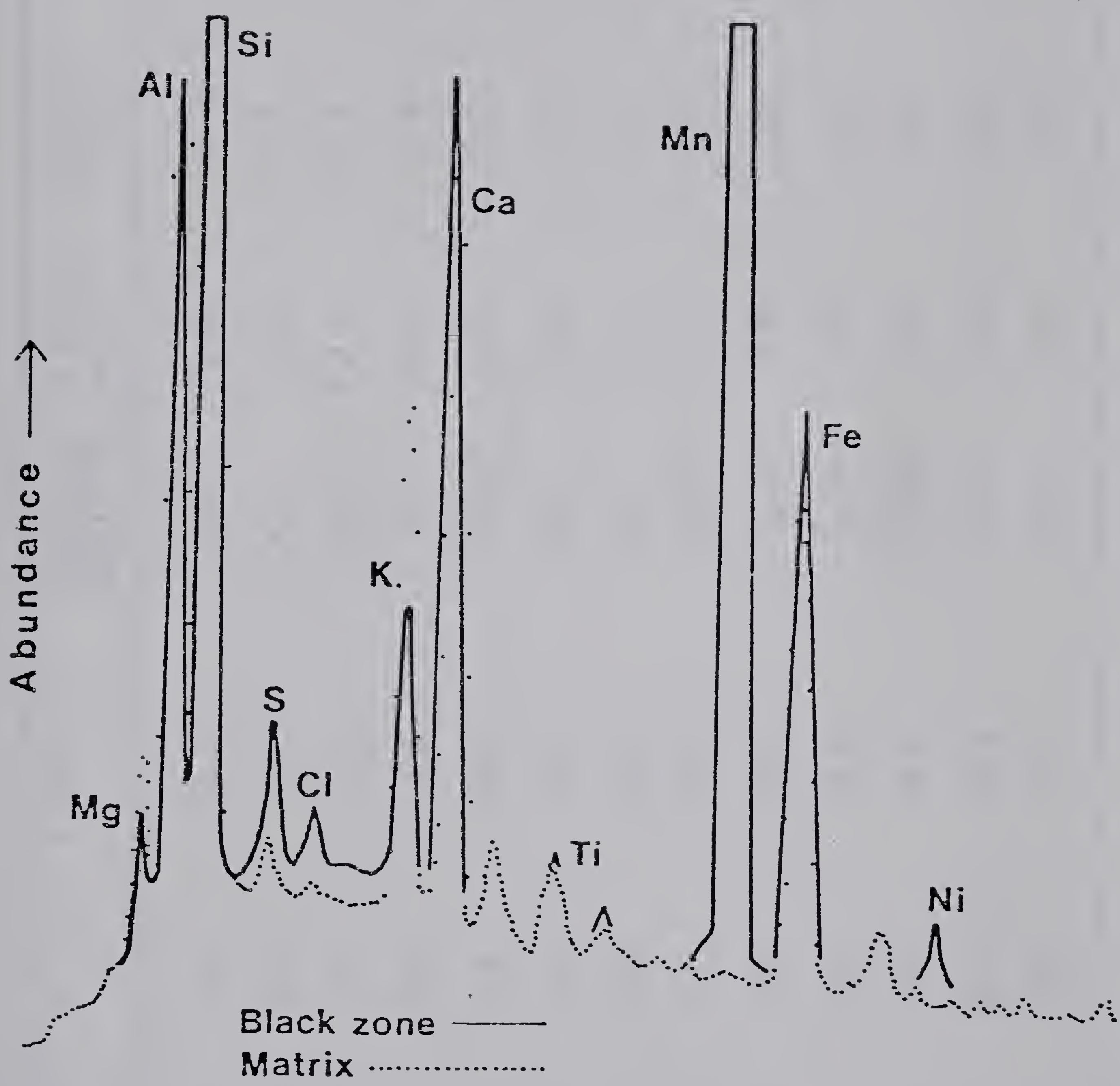


Figure 3. C horizon joint plane surface: energy dispersive X-ray analysis.

Table 6 SELECTED PHYSICAL AND CHEMICAL PROPERTIES

Horizon	% Sand	% Silt	% Clay	% Fine Clay	$\frac{\text{Fine Clay}}{\text{Total Clay}}$	% $CaCO_3$ equiv.	% Organic Carbon	% N	$\frac{C}{N}$	pH
L-F	n.d.*	n.d.	n.d.	n.d.	n.d.	n.d.	48.5	1.80	26.9	7.2
F-H	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	39.4	2.20	17.9	6.9
H	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	41.1	2.40	17.1	6.6
Ah1	8.5	43.4	48.1	19.9	0.41	n.d.	5.9	0.53	11.1	6.1
Ah2	9.3	45.1	45.6	16.4	0.36	n.d.	2.7	n.d.	n.d.	6.4
Ah3	10.2	43.5	46.3	17.3	0.37	n.d.	1.4	0.16	8.8	6.8
Ahej	8.6	42.8	48.6	23.3	0.48	n.d.	0.5	0.08	6.3	6.7
$\overline{\text{Aegj}}$	11.0	47.5	41.6	18.6	0.45	n.d.	0.3	0.06	5.0	6.9
Bmgj	6.7	40.4	52.9	18.7	0.35	0.00	0.5	0.07	7.1	7.3
Bmkgj	16.1	33.2	50.6	14.8	0.29	0.48	0.4	0.06	6.7	7.3
BCkgj	4.5	34.5	61.0	18.8	0.31	0.52	0.6	0.07	8.6	7.6
Ckgj1	2.8	34.0	63.2	18.3	0.29	1.10	0.6	0.06	10.0	7.8
Ckgj2	1.6	34.7	63.7	18.5	0.29	1.04	0.6	0.06	10.0	7.4

* n.d. - not determined

narrowed abruptly between the organic and Ah horizons and reached their lowest values in the Ahej to Bmgj horizons.

5.2.4 Carbonates, pH, and Cation Exchange

Carbonates were present in amounts of approximately 1% or less, with the highest content in the Ckgj horizon (Table 6).

Soil reaction values were in a narrow range, from slightly acid to mildly alkaline (pH 6.1 to 7.8) (Table 6). The lowest value was obtained for the Ah1 horizon, from which the pH rose both toward the surface and the base of the pedon.

Cation exchange capacity values showed good agreement between the two methods used: determination directly and by summation of the exchangeable Ca^{++}, Mg^{++}, Na^{+}, and K^{+} (Table 7). This indicated essentially complete base saturation. The highest C.E.C. values were found in the organic horizons, increasing from the L-F to H. In the mineral horizons, C.E.C. peaked in the Ah1, reflecting its high organic matter content, and in the C horizons, probably as a result of their high clay content.

The dominant exchangeable cation was Ca^{++}, with lesser amounts of Mg^{++}, K^{+}, and Na^{+}, in that order. Exchangeable Ca^{++} was most abundant in the H layer, with a secondary maximum in the C horizons, while Mg^{++} peaked in the Ahej horizon. Within the organic horizons, K^{+} decreased in abundance from the L-F to the H horizon (3.2 to 1.2%), while Na^{+} displayed a similar trend (0.4 to 0.1%). However, within the mineral horizons, exchangeable K^{+} peaked at the uppermost horizon (2.4%) and declined to a minimum of 1.0% in the Ckgj2 horizon, while exchangeable Na^{+} increased from 0.1 to 1.0%.

5.2.5 Extractable Iron and Aluminum

Table 7 CATION EXCHANGE PROPERTIES

Horizon	C.E.C. (m.e./100 g)		% of Exchangeable Cations			
	Det.	Sum	Ca^{++}	Mg^{++}	K^{+}	Na^{+}
L-F	138.6	133.6	70.4	22.5	3.2	0.4
F-H	140.7	139.9	74.6	21.7	2.6	0.4
H	159.9	155.9	76.2	20.0	1.2	0.1
Ah1	54.2	54.2	68.6	28.8	2.4	0.1
Ah2	n.d[1]	n.d	n.d	n.d	n.d	n.d
Ah3	40.1	39.9	61.3	36.2	1.7	0.2
Ahej	35.9	35.9	58.8	39.0	1.9	0.3
Aegj	24.1	23.5	60.6	34.9	1.7	0.4
Bmgj	39.1	37.0	60.1	32.5	1.5	0.3
Bmkgj[2]	43.0	42.1	72.3	24.0	1.4	0.5
BCkgj[2]	55.0	53.7	72.7	23.1	1.1	0.5
Ckgj1[2]	51.1	50.1	73.4	22.1	1.2	1.2
Ckgj2[2]	59.4	59.4	73.9	24.1	1.0	1.0

[1] n.d. - not determined

[2] Carbonates present; values include both exchangeable and extractable cations.

For Fe, each of the three extraction methods gave a different pattern of variation within the pedon (Table 8). While dithionite-extractable Fe (Fe_d) fluctuated within a narrow range from the Ahej to the Ckgj2 horizon, 0.87 to 1.17%, its concentration was approximately half of those values in the Ah horizons. By contrast, oxalate-extractable Fe (Fe_o) was less abundant throughout the profile and showed much less of a range of concentrations (0.21 to 0.32%). Pyrophosphate-extractable Fe (Fe_p) was somewhat more abundant than Fe_o in the Ah1 horizon, but its concentration declined greatly with depth, roughly parallelling the trend in organic matter content.

Extractable Al contents varied in a different fashion from Fe. Al_d and Al_o were highest in the Ah1 to Bmgj horizons. Al_p also peaked in the Ah1 but its lower concentrations in the C horizons did not represent as large a decrease as for Fe_p.

5.2.6 Fine Sand Composition

Assigning all of the Na and Ca in the < 2.90 sp gr fraction of the 50 to 250 μm sand to feldspars, it appeared that the sodic forms were overwhelmingly more abundant (Table 9). The proportions of sodic vs calcic feldspars, as inferred from the Na: Ca ratios, showed no systematic variation that had any obvious relationship to depth or genetic horizon. Likewise, the content of heavy minerals (> 2.90 sp gr) showed an irregular pattern of variation, with wide fluctuations occurring between adjacent horizons.

5.2.7 Clay Mineralogy

Variations in mineralogy were greater between the two size fractions

Table 8 EXTRACTABLE IRON AND ALUMINUM

Horizon	Fe_p [1]	Fe_o [2]	Fe_d [3]	Al_p [1]	Al_o [2]	Al_d [3]
Ah1	0.37	0.31	0.44	0.34	0.20	0.09
Ah2	0.14	0.21	0.39	0.16	0.19	0.08
Ah3	0.07	0.21	0.53	0.06	0.15	0.09
Ahej	0.11	0.23	0.87	0.13	0.13	0.10
A$\overline{\text{egj}}$	0.13	0.26	0.90	0.14	0.12	0.09
Bmgj	0.09	0.27	1.14	0.10	0.13	0.10
Bmkgj	0.06	0.26	0.99	0.09	0.11	0.08
BCkgj	0.06	0.32	1.07	0.08	0.10	0.06
Ckgj1	0.07	0.32	1.12	0.10	0.09	0.06
Ckgj2	0.07	0.31	1.17	0.10	0.10	0.07

1 Na-pyrophosphate extraction

2 Acid ammonium oxalate extraction

3 Dithionite - citrate - bicarbonate extraction

Table 9 FINE SAND (50-250 μm) COMPOSITION

Horizon	< 2.93 sp gr fraction % Soda-calcic feldspar	Na:Ca	% Heavy Minerals (> 2.93 sp gr)
Ah1	13.8	66.7	1.15
Ah2	10.8	37.2	0.91
Ah3	10.7	51.7	0.95
Ahej	10.6	48.4	1.39
Aegj	10.7	58.1	1.09
Bmgj	12.0	65.0	0.99
Bmkgj	18.9	38.8	1.11
Bckgj	8.4	38.4	1.29
Ckgj1	10.1	51.8	1.37
Ckgj2	9.6	22.2	2.22

(< 0.2 μm and 0.2-2.0 μm) than between soil horizons (Table 10). As a preliminary generalization, the fine clays were dominated by smectite and mica, while the coarse clays had in addition kaolinite, chlorite, and small amounts of quartz (Figure 4, Table 10).

In the coarse clay fraction, smectite accounted for about 35-40% with the C.E.C. method usually providing a higher estimate. Mica content showed a similarly narrow range, 30 to 33.5%, while kaolinite, as estimated from X-ray diffraction peak intensity, probably accounted for about 10 to 15%. The balance of the crystalline components consisted of chlorite and quartz, with no feldspar diffraction peaks detected. There was remarkably little variation between horizons, with the differences observed probably within the range of error of the methods employed. The content of amorphous components was limited, as evidenced by sharp diffraction peaks despite the lack of pretreatments for organic matter or sesquioxide removal.

There were only two features which complicated this simple coarse clay mineralogy. A small amount of a probable Al-hydroxy intergrade mineral apparently occurred in the Ah2 horizon, as shown by a poorly defined peak in the 12-13 Å region for the K-saturated, 0% relative humidity treatment; this peak collapsed with heating to 300°C and 550°C (Figure 4). Secondly, since Abder-Ruhman (1980) detected appreciable quantities of vermiculite in soils of the adjacent east-central region of Alberta, it is of interest that similar criteria suggested the presence of this mineral at the study site. For the pretreatments employed, vermiculite does not give uniquely distinctive X-ray diffraction peaks, but it coincides with those of smectite and chlorite. However, for K-saturated samples, vermiculite does not rehydrate after the 0% relative humidity treatment. Therefore, one can compare the relative intensity ratios of 10 Å : 7 Å peaks

Table 10

CLAY MINERALOGY

Horizon	% Smectite[1] Fine Clay	% Smectite[1] Coarse Clay	% Smectite[2] Fine Clay	% Smectite[2] Coarse Clay	% Smectite (Mean) Fine Clay	% Smectite (Mean) Coarse Clay	% Mica[3] Fine Clay	% Mica[3] Coarse Clay	% Kaolinite[4] Fine Clay	% Kaolinite[4] Coarse Clay	% Chlorite[4] Fine Clay	% Chlorite[4] Coarse Clay	% Quartz[4] Fine Clay	% Quartz[4] Coarse Clay
Ah1	54.6	29.4	83.4	54.5	69.0	42.0	22.5	32.5	<5%	10-15	n.d.[5]	5-9	n.d.	3-6
Ah2	62.0	26.7	72.1	47.0	67.1	36.9	21.5	30.7	<5%	10-15	n.d.	5-9	n.d.	3-6
Ah3	61.2	32.7	71.7	50.6	66.5	41.7	22.3	31.1	<5%	10-15	n.d.	4-7	n.d.	n.d.
Ahej	64.1	29.6	66.2	39.5	65.2	34.6	22.6	33.5	<5%	10-15	n.d.	4-7	n.d.	2-4
Aegj	66.7	35.8	64.4	40.4	65.6	38.1	22.3	31.2	<5%	10-15	n.d.	5-9	n.d.	2-4
Bmgj	65.4	35.2	63.5	40.8	64.5	38.0	24.2	30.6	<5%	10-15	n.d.	5-9	n.d.	2-4
Bmkgj	64.9	35.4	66.0	47.7	65.5	41.6	23.5	30.0	<5%	10-15	n.d.	6-10	n.d.	3-6
BCkgj	68.4	30.3	71.1	45.8	69.8	38.1	25.3	30.3	<5%	10-15	n.d.	5-9	n.d.	2-4
Ckgj1	66.0	29.1	59.0	44.5	62.5	36.8	25.0	31.2	<5%	10-15	n.d.	5-9	n.d.	2-4
Ckgj2	63.4	32.9	59.3	50.0	61.4	41.5	27.4	32.1	<5%	10-15	n.d.	5-9	n.d.	2-4

1 estimated from surface area, using 850 m^2/g for montmorillonite standard

2 estimated from cation exchange capacity, using 110 m.e./100 g for smectite

3 based on 10% K_2O content

4 estimated from X-ray diffraction patterns

5 n.d. - not detected

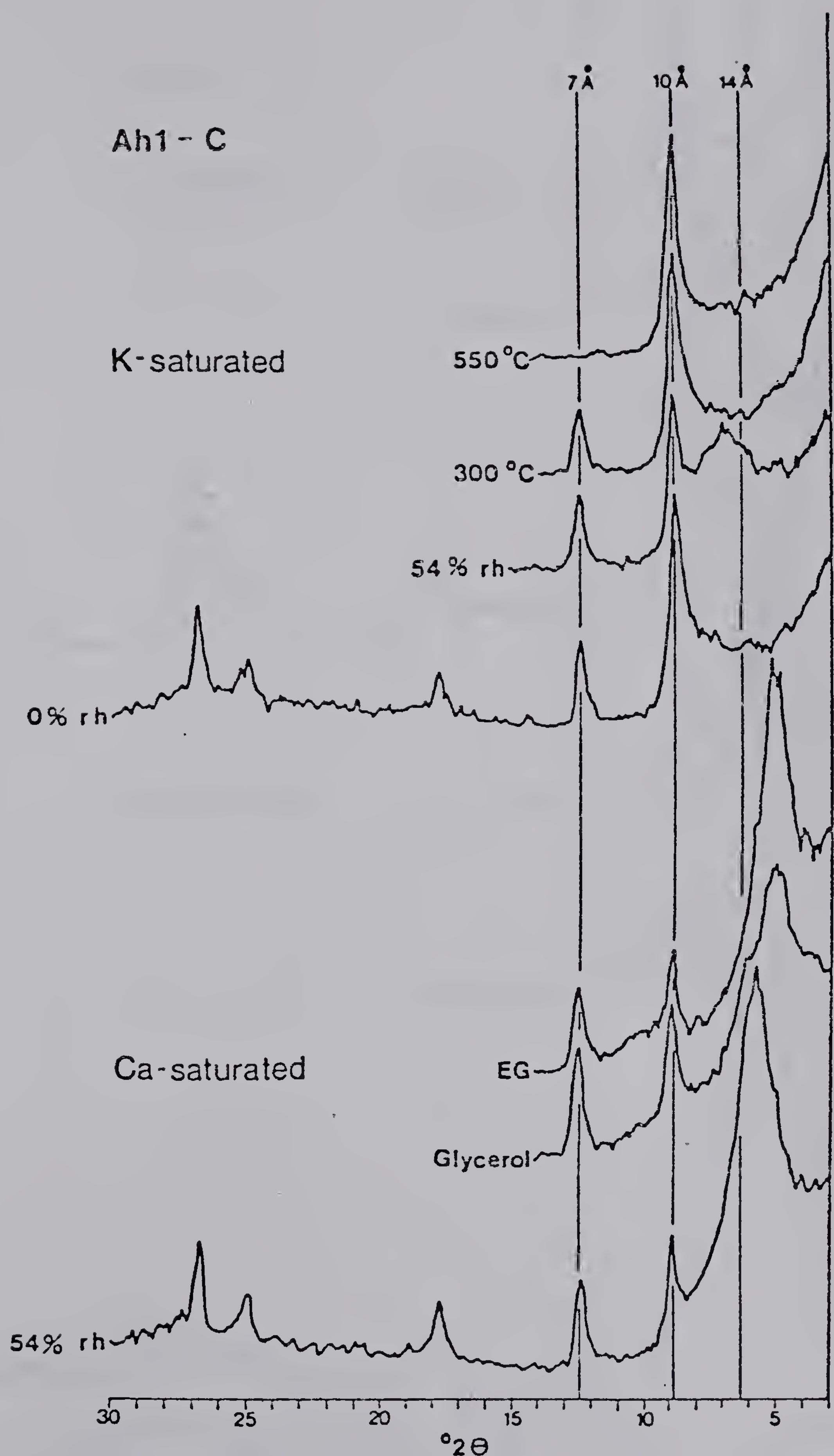
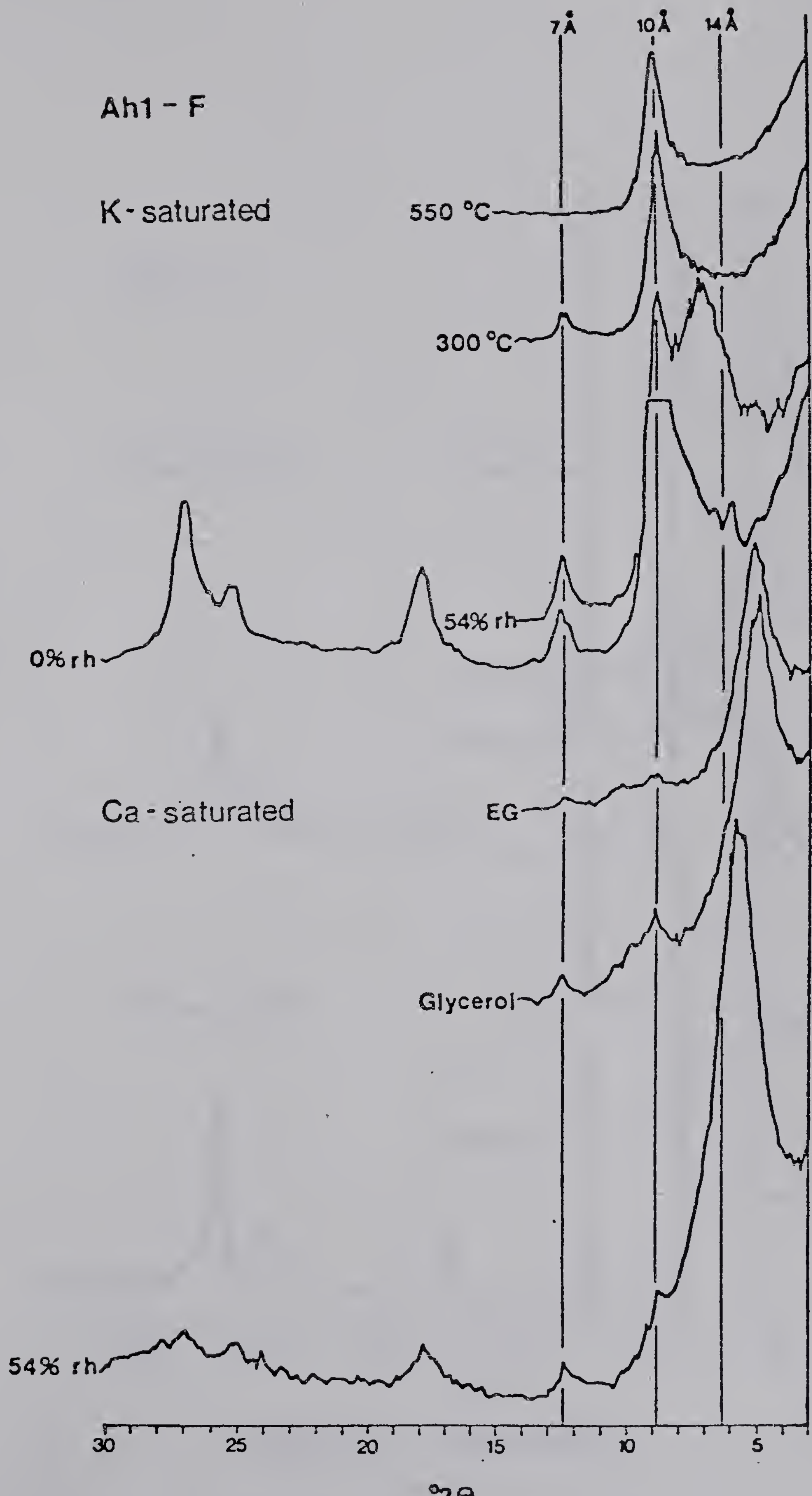


Figure 4. X-ray diffractograms for selected clay fractions. (C - 0.2-2.0 μm clay, F - <0.2 μm clay, rh - relative humidity, EG - ethylene glycol)



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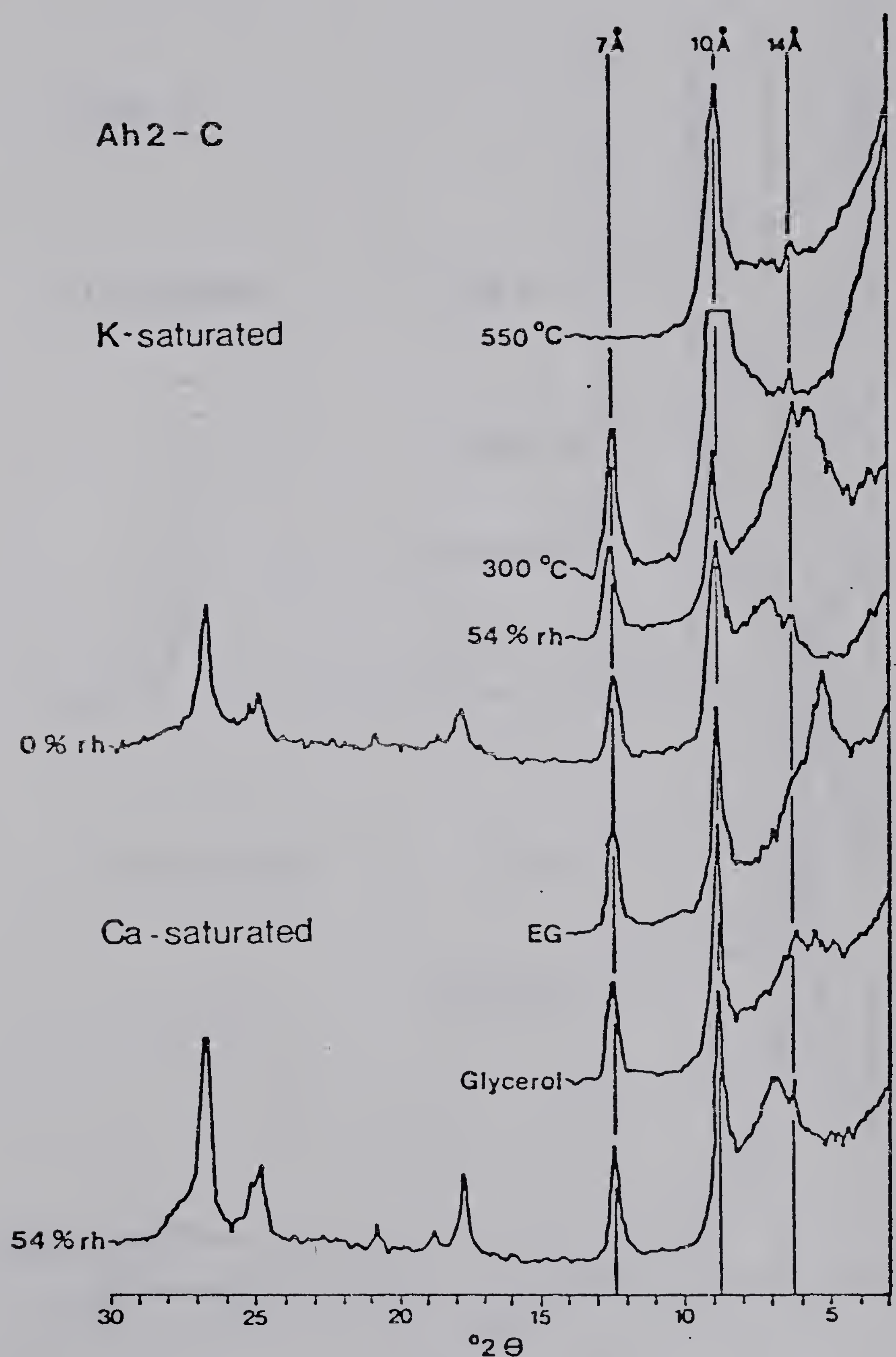

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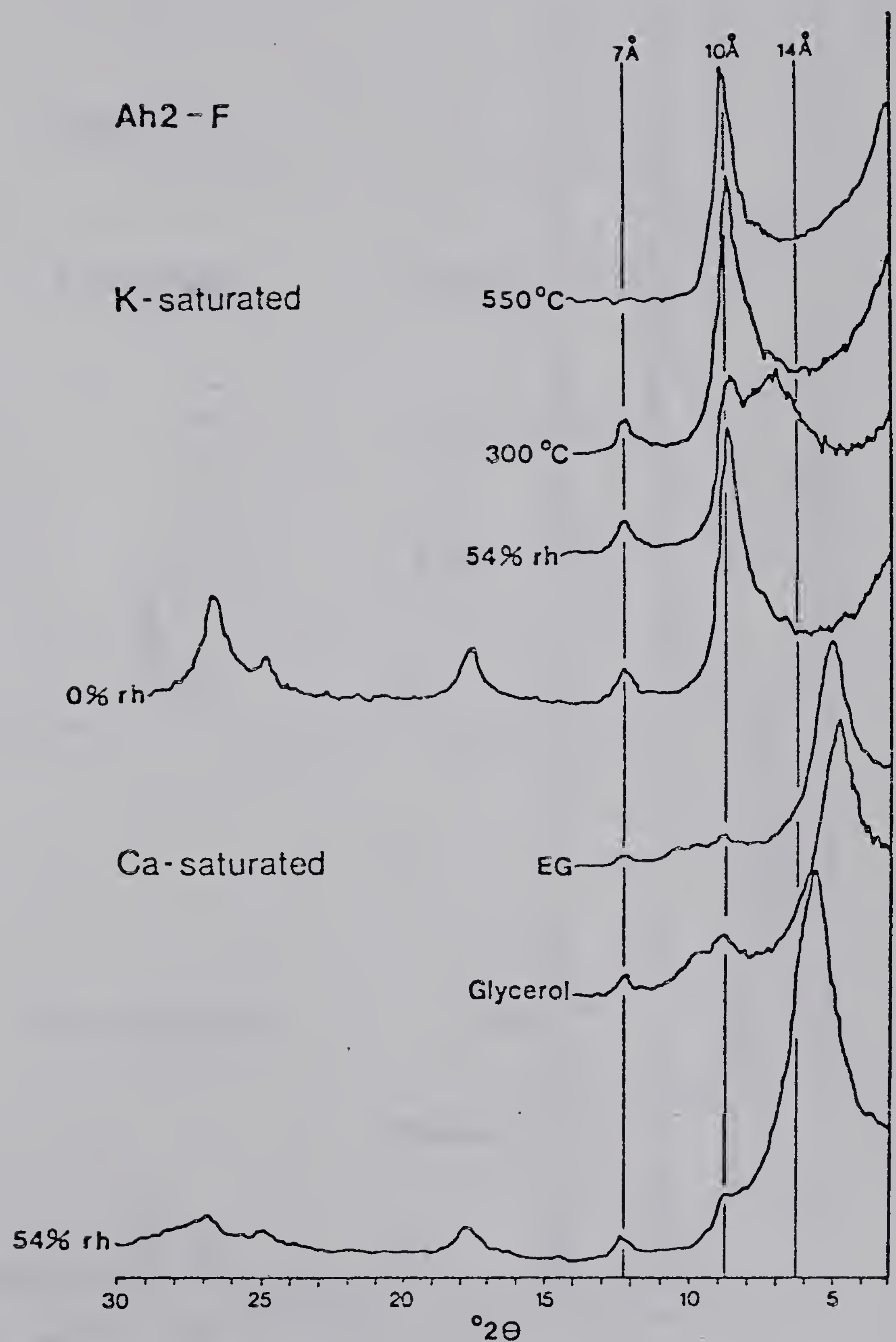

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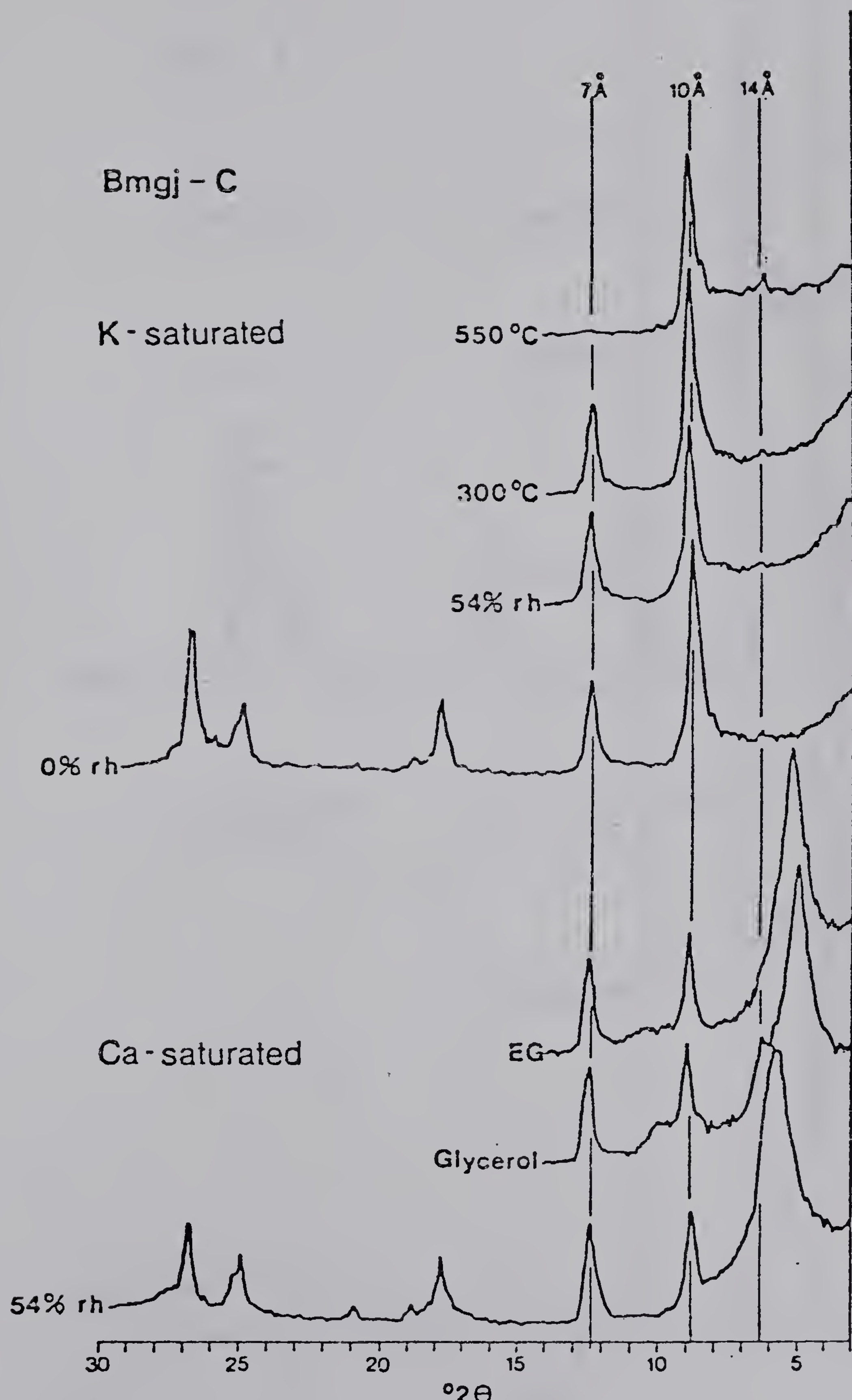

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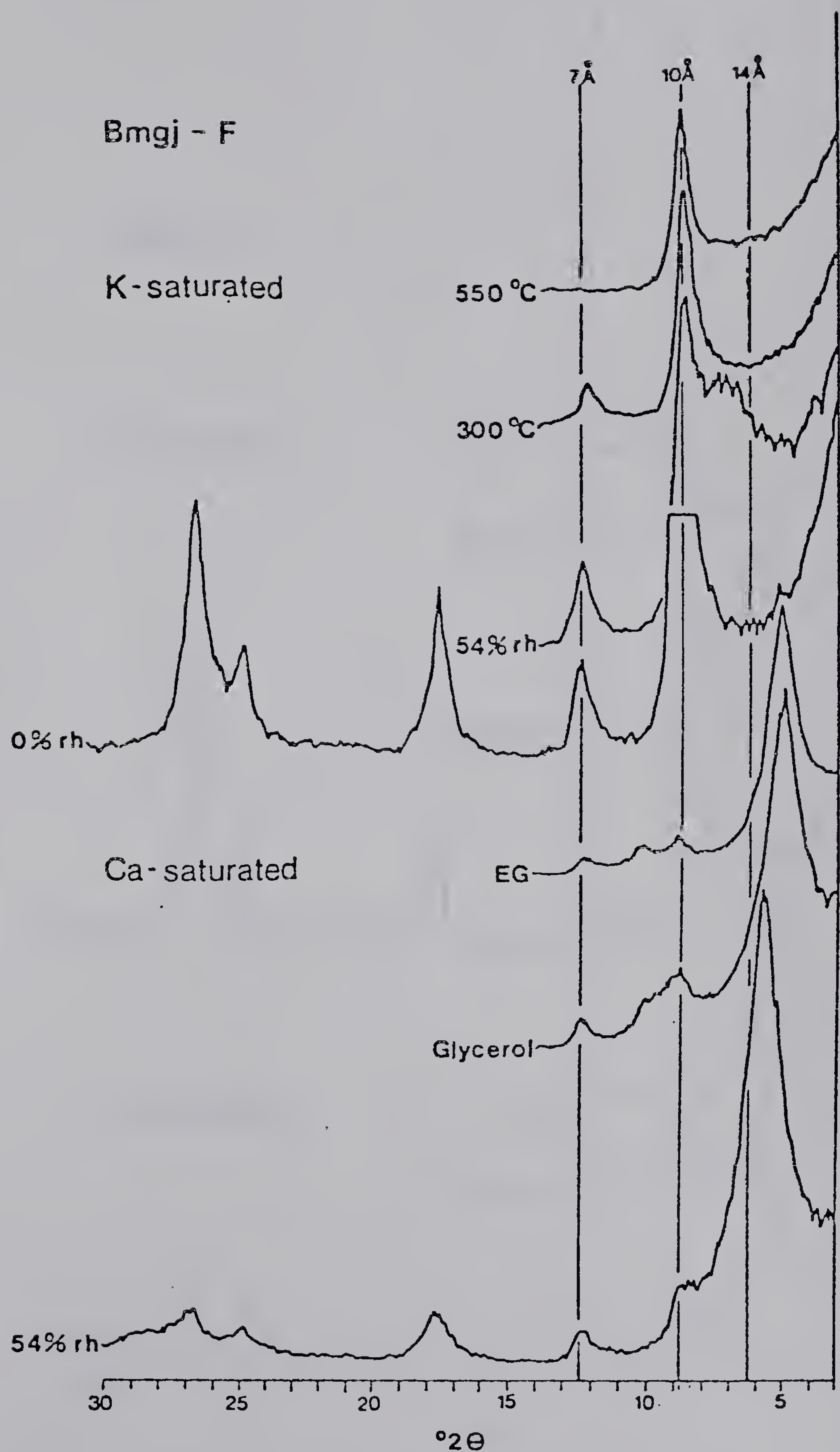

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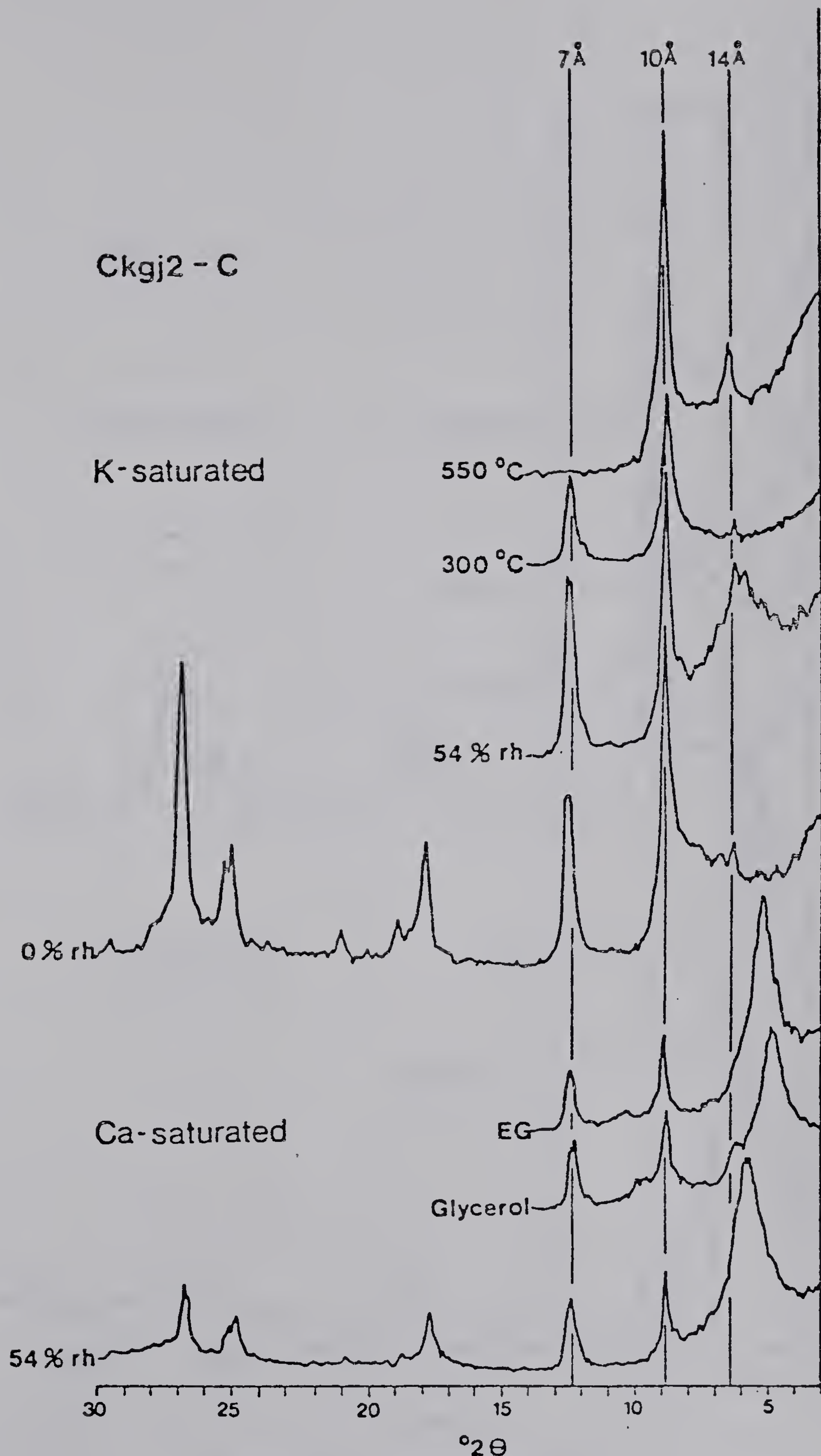

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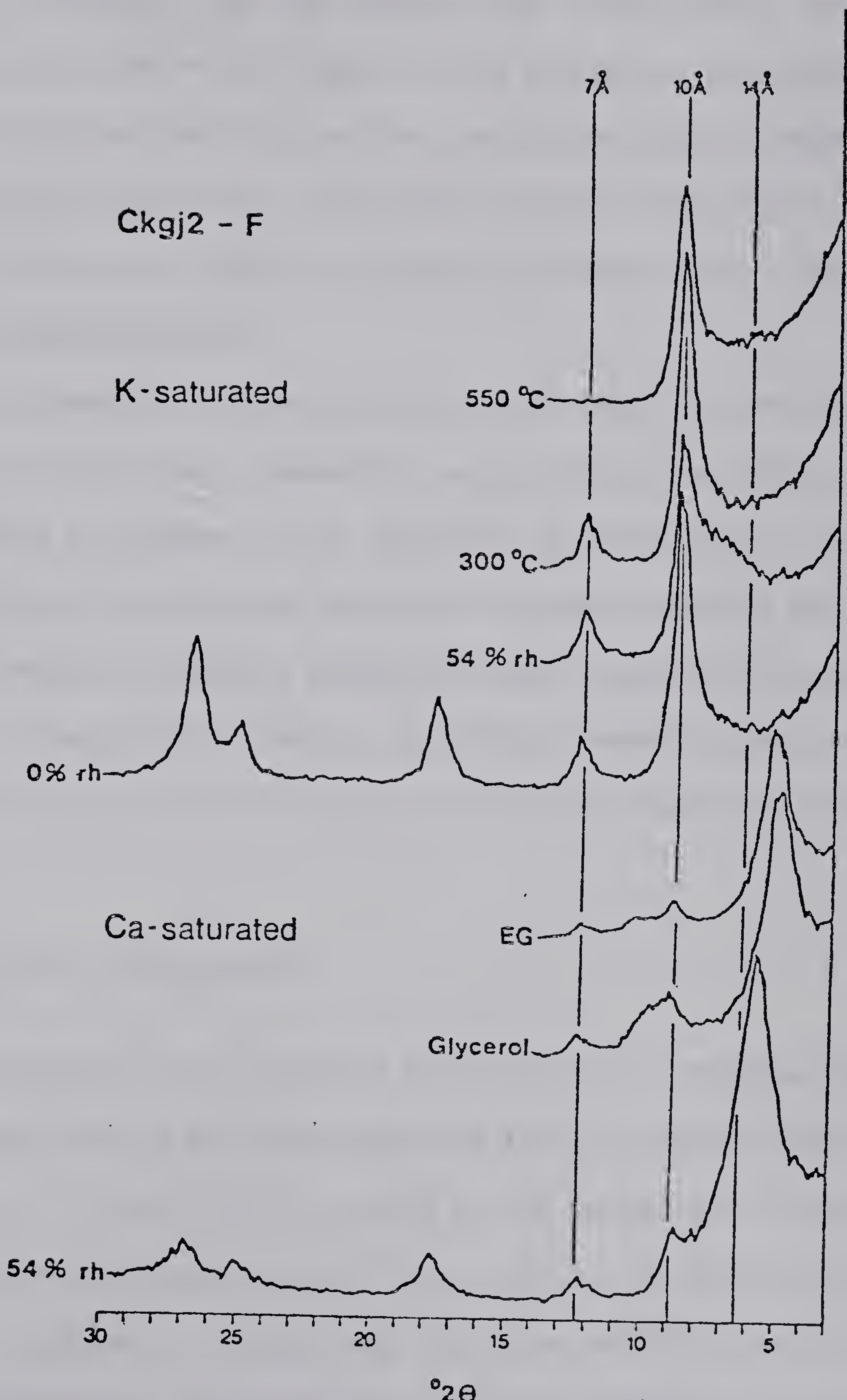

Figure 4. (Continued)

in Ca- and K-saturated samples at 54% relative humidity, with higher ratios for the K-saturated clay indicating that a non-rehydrating mineral, likely vermiculite, is present. For the coarse clays in this pedon, these ratios were virtually the same in the Bmkgj horizon and below, but were usually higher at 54% relative humidity for the K-saturated than Ca-saturated samples in the upper horizons. Thus some portion of the high C.E.C. and surface area values was likely attributable to vermiculite, albeit a small amount, in the upper horizons.

Fine clay mineralogy was much simpler, with about 75% smectite, 20-25% mica, and the balance apparently consisting of kaolinite. Using the same criteria as before, it was difficult to determine if vermiculite was present, since the prominent peak from hydrated smectite for the Ca-saturated 54% relative humidity treatment almost completely overwhelmed the 10 Å peak. Tentatively, though, the effect seemed to have occurred, suggesting the presence of small amounts of vermiculite in the fine clay fraction.

5.2.8 Clay-Organic Complexation

Carbon contents of the untreated Ah horizon clay fractions decreased with depth, parallelling the trend observed for the organic matter content of the whole soil. From 7.29% and 3.41% in the coarse and fine clays of the Ah1, C content decreased to 3.05% and 1.19% in the Ah3 (Table 13). In the fine clay fractions, C content was less than half that of the coarse clay. This difference was expressed in the much darker colour of the coarse clay separate. The difference in C content between the fine and coarse clays was much greater at this site than for the Orthic Black Chernozems studied by Dudas (1968). Significantly, the northernmost pedons

(Edmonton area) examined in that study had the relative fine vs. coarse clay C content ratios closest to those from the Ellerslie site.

For the Ah2 and Ah3 horizons, the total C.E.C. values for the untreated fine clays were higher than for the untreated coarse clays: 95.3 vs. 86.8 m.e./100 g (Ah2) and 87.2 vs. 76.3 m.e./100 g Ah3 (Table 11). This difference reflects the dominance of smectite in the fine clay fraction. However, in the Ah1, the untreated coarse clay had a higher total C.E.C. than the fine clay (112.2 vs. 98.9 m.e./100 g), probably as a result of the high organic matter content of the former.

The pH-independent C.E.C. of the untreated clays was higher for the fine clay fraction than the coarse clays, presumably reflecting their higher smectite content. Values ranged from 66.3 to 69.6 m.e./100 g for the fine clays and from 48.6 to 66.3 m.e./100 g for the coarse clays. Removal of organic matter usually decreased the pH-independent C.E.C., a phenomenon attributed by Dudas and Pawluk (1969a) to the high concentration of KCl possibly exchanging with H^+ from organic functional groups. In addition, these authors also suggested that organic matter could be responsible for physical chemical adsorption of K^+.

The pH-dependent C.E.C. of the untreated clays decreased with depth and particle size, a trend parallel to that of carbon content. Hydrogen peroxide-treated samples showed higher pH-dependent C.E.C. values in the fine clays than in the coarse clays, an effect attributed to a greater number of exposed clay hydroxyl groups (Dudas and Pawluk, 1969a).

The three types of C.E.C. for the six samples showed a decrease in the majority of cases after H_2O_2 treatment (Table 12). For the total C.E.C., the magnitude of the decrease between untreated and treated samples was reduced as particle size decreased. While total carbon content

Table 11 CATION EXCHANGE CAPACITIES OF Ah HORIZON CLAY FRACTIONS*

Horizon	Fraction	Total C.E.C.		pH-indep. C.E.C.		pH-dep. C.E.C.	
		Untreated	H_2O_2-treated	Untreated	H_2O_2-treated	Untreated	H_2O_2-treated
Ah1	0.2-2.0 μm	112.2	95.3	66.3	35.4	45.9	59.9
	<0.2 μm	98.9	113.8	69.6	67.2	29.3	46.6
Ah2	0.2-2.0 μm	86.8	72.8	57.1	31.9	29.7	40.9
	<0.2 μm	95.3	85.7	66.3	65.5	29.0	20.2
Ah3	0.2-2.0 μm	76.3	61.1	48.6	38.7	27.7	22.4
	<0.2 μm	87.2	87.9	66.3	71.6	20.9	16.3

* m.e./100 g clay

Table 12 EFFECT OF H_2O_2 TREATMENT ON C.E.C. OF Ah HORIZON CLAY FRACTIONS*

Horizon	Fraction	Total C.E.C.	pH-indep. C.E.C.	pH-dep. C.E.C.
Ah1	0.2-2.0 μm	-16.9	-30.9	+14.0
	<0.2 μm	+14.9	-2.4	+17.3
Ah2	0.2-2.0 μm	-14.0	-25.2	+11.2
	<0.2 μm	-9.6	-0.8	-8.8
Ah3	0.2-2.0 μm	-15.2	-9.9	-5.3
	<0.2 μm	+0.7	+5.3	-4.6

* m.e./100 g clay

Table 13 CARBON CONTENT OF Ah HORIZON CLAY FRACTIONS

Horizon	Fraction	% C
Ah1	0.2-2.0 μm	7.29
	<0.2 μm	3.41
Ah2	0.2-2.0 μm	5.09
	<0.2 μm	1.75
Ah3	0.2-2.0 μm	3.05
	<0.2 μm	1.19

also decreased with particle size, it is significant that the decrease in total C.E.C. after peroxide treatments, particularly for the fine clays, was less than would be expected as a result of organic matter removal. In fact, total C.E.C. of the Ah1 and Ah3 horizon fine clays actually increased after peroxide treatment.

5.2.9 Humic Acid Infrared Spectra

There were only subtle differences in the infrared spectra of extracted humic acids for the L-F, H, Ah1, Ahej, and Bmgj horizons. Spectral characteristics are only listed here; their significance will be discussed later.

For the 2500-1800 cm^{-1} spectral region, a slight upward slope was noted for the three upper horizons, while for the Ahej and Bmgj humic acids the same region was essentially flat (Figure 5). Three other absorption bands identified in humic acids by Dormaar (1967) and Dormaar and Lutwick (1966) appeared in these samples as follows. The 2920 and 2850 cm^{-1} bands indicate aliphatic CH_x groups; the former was strongest in the L-F humic acid spectrum and absent or extremely weak in the others. Carbonyl stretch of carboxylic acids or of ester linkages is indicated by absorption at 1725 cm^{-1}. This band was strongest in the L-F and H horizon spectra and was absent lower in the solum. Finally, the 1525 cm^{-1} band, which is assigned to aromatic C=C, was not observed in any of the spectra.

5.2.10 Discussion

While the Ah horizons of this pedon meet the requirements of the Chernozemic A horizon (Appendix) and the colour criteria of the Black Chernozemic Great Group, the presence of prominent organic horizons is

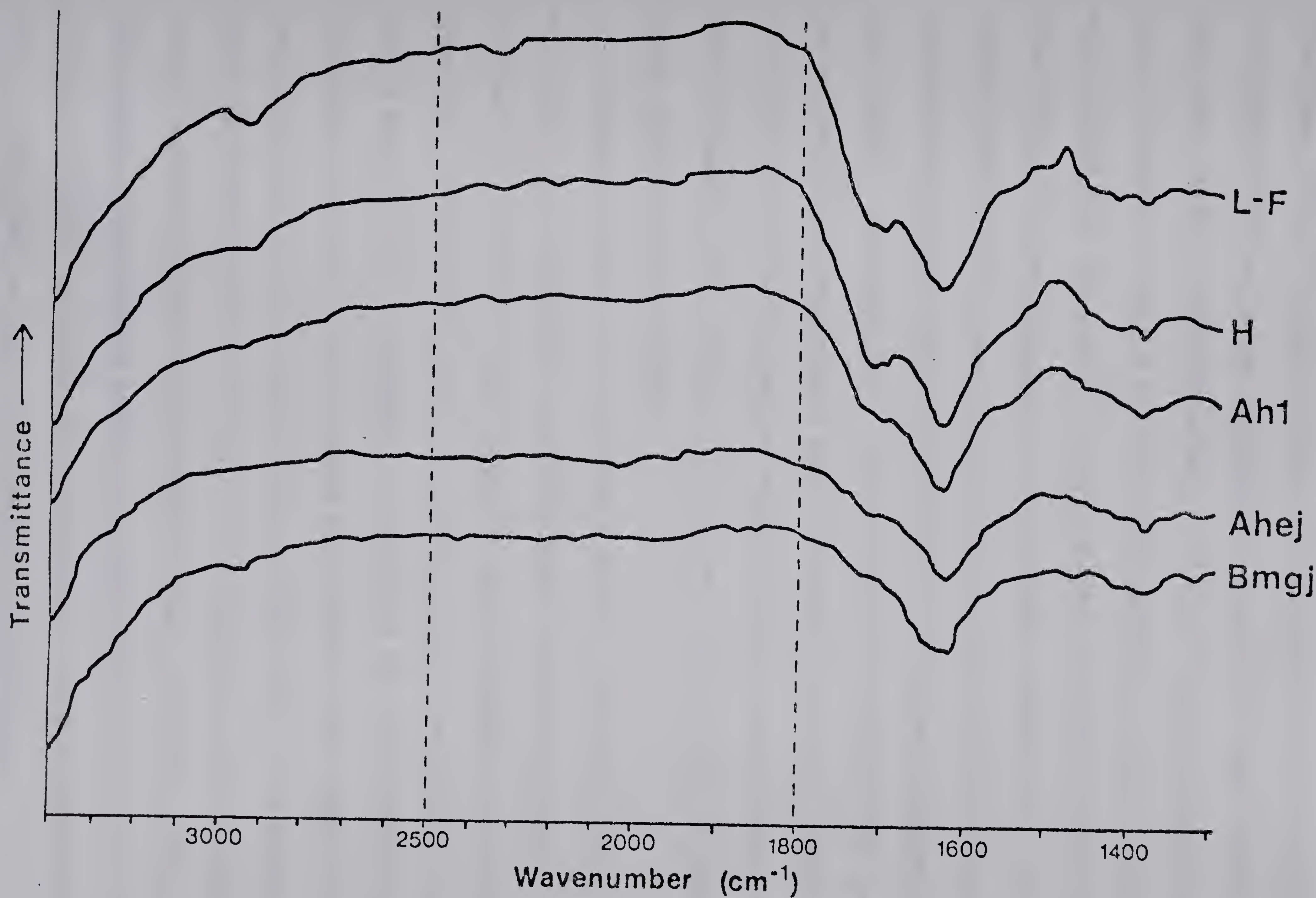

Figure 5. Infrared spectra of humic acids from selected horizons.

more characteristic of the Dark Gray Great Group (Canada Soil Survey Committee, 1978). As noted in the earlier discussion of micromorphology, the organic horizons showed considerable evidence of faunal activity, with the H horizon consisting almost entirely of fecal pellets. The great thickness of the F-H and H horizons compared to that of the L-F may indicate that the rate of litter breakdown is comparatively rapid; otherwise, litter would accumulate to a greater thickness (cf. Bal, 1970). Although there was an abrupt boundary between the H and Ah1 horizons, thin sections revealed that some mineral material had become incorporated in the H horizon structural units. This suggests that some of the fauna present may ingest both mineral soil and organic detritus. However, the existence of a discrete H horizon implies that the rate of mixing of humified material with mineral soil is slower than the rate of organic matter addition at the surface.

Compared to those in the F-H and H horizons, the aggregates in the Ah horizons were both larger and much richer in mineral material, suggesting that other factors or other organisms may be responsible for their formation. Traditionally, mull formation has been attributed in large part to the activity of earthworms in ingesting and mixing both organic and mineral matter and leaving behind fecal pellets that comprise the dominant structural units (Babel, 1975; Kubiena, 1953). The tendency toward uniformity of shape and size in such mullgranic units has been cited as evidence of their faunal origin (Brewer and Pawluk, 1975). While other mesofauna may perform similar structure-promoting functions, the question of whether earthworms are indigenous to the Black Chernozemic soils of North America seems to be a matter of controversy.

For example, Buntley and Papendick (1960) found that earthworms

appeared to be recent invaders of Chernozems in the eastern U.S. Great Plains, as shown by almost complete destruction of the normally abrupt horizon boundaries. The implication would seem to be that the Black Chernozemic soils of this continent do not owe their morphology to earthworm activity, unlike their European counterparts (cf. Kubiena, 1953). The possible role of other mesofauna has been given little attention by researchers, with Ah formation in these soils usually attributed to in situ decomposition and humification of root residues. However, the literature is usually vague as to how this would produce the typical granular structure of such horizons.

The thickness and organic carbon and nitrogen contents of the Chernozemic A horizon in this pedon were comparable to the values reported for other pedons in the Black soil zone of western Canada (Dudas and Pawluk, 1969b; St. Arnaud and Whiteside, 1963) and were close to those deemed typical of the Malmo series (Bowser et al., 1962).

The cation exchange capacity determinations for the Ah horizon clay fractions demonstrated the close association of organic and mineral materials in the fine clay fraction. The fact that total C.E.C. increased or decreased only slightly after organic matter was removed from the fine clay fractions indicated that functional groups blocked by or involved in clay-organic complexation had been freed. Dudas and Pawluk (1969a) suggested that similar results from other Black Chernozemic soils in Alberta likely reflected two factors: (1) in untreated clays, some of the organic functional groups were not available for exchange reactions, and (2) clay hydroxyl groups were involved in complexation with organic matter and could only contribute to the C.E.C. after peroxide treatment. As in that previous study, the results for the Ellerslie pedon suggest

that the degree of clay-organic complexation is greatest in the fine clay fraction, despite the lower carbon content.

Interpretation of the humic acid infrared spectra is of interest to this study because the work of Dormaar (1967) suggested a relationship between certain spectral characteristics and the vegetation history of a site. In particular, the slope of the spectral line in the 2500-1800 cm^{-1} region was found to vary according to whether the dominant vegetation had been grassland or forest. Chernozemic humic acids developed under grassland showed an upward slope in that region, while humic acids from Podzolic and Luvisolic soils developed under forest had a downward slope. Dormaar (1967) claimed that this effect was based on the average molecular weight of humic acids being smaller in the latter soil groups. By way of demonstration, when humic acid from a grassland soil was separated into fractions of progressively lower molecular weight, the spectra gradually changed to a more "Podzolic" character. This relationship was compared to the situation in forest-invaded Chernozemic soils in which mobile, lower molecular weight humic substances, characteristic of Luvisols, would begin to accumulate in the B horizon. However, this interpretation must be tempered by conflicting evidence, as shown by Dawson et al. (1978). In that study, spectra for low molecular weight (837-845) mobile fulvic acids from a Podzol lysimeter leachate displayed essentially no slope in the 2500-1800 cm^{-1} region.

A later study by Dormaar (1973) pointed out another complication, in that a flattening of the same spectral region occurred in spectra of humic acids from Bg horizons of Humic Gleysols. This pattern was attributed to the lack of removal of low molecular weight condensation byproducts of humic acid formation under poorly drained conditions.

The three other bands discussed by Dormaar (1967) were expected to

intensify with an increasingly "Podzolic" character of the humic acids. These bands are taken to indicate a lower degree of condensation (2920, 2850 cm^{-1}), greater acidity (1725 cm^{-1}), and a higher content of aromatic structures (1525 cm^{-1}).

Given these criteria, the humic acid infrared spectral characteristics noted earlier for this pedon do not lend themselves to an unambiguous interpretation. For example, the slopes of the 2500-1800 cm^{-1} region in the mineral soil humic acids could indicate either hydromorphic influences or the early stages of transformation to a Luvisolic soil. Since other evidence has suggested that the pedon is at best only moderately well drained, the first interpretation cannot be dismissed. Furthermore, the three other spectral bands, whose intensification is thought to indicate a "Podzolic" character, are either absent or weakly expressed in the mineral soil humic acid spectra.

The low carbonate content of the glaciolacustrine parent material is similar to values reported from elsewhere in the Lake Edmonton Plain (Arshad and Pawluk, 1966; Bowser et al., 1962) and is lower than that typical of glacial tills in the same region (Dudas, 1968; Abder-Ruhman, 1980). Despite this lower lime content in the parent material, the near-neutral pH values and the high base saturation are typical of Black Chernozemic soils elsewhere in the Prairie region. The dominance of exchangeable Ca^{++}, that reaches its maximum level in the upper horizons, probably reflects a high rate of return in the litterfall. The fact that exchangeable K^{+} achieved peak abundance in the upper horizons as well, also suggests the influence of nutrient cycling, although its release through the weathering of micaceous minerals is also a possiblity. The increase in exchangeable Na^{+} with depth likely indicates both the greater mobility of this

element and the fact that parent materials of the Malmo series may sometime be saline (Bowser et al., 1962).

The stratified, texturally variable nature of the soil parent material is typical of the Lake Edmonton sediments and presents difficulties in interpreting certain aspects of pedogenesis. First, one of the criteria for determining if lessivage has occurred is an enrichment of clay, particularly fine clay, in the B horizon compared to the A horizon. In fact, as shown by the fine clay: total clay ratios, the opposite was the case. While this could simply be an inherited property overshadowing the effects of illuviation, there was also no evidence for illuvial argillans in the B horizons. Rare void argillans were found in the Ckgj2 horizon, but these were likely stress features related to shrink-swell processes. Thus, the evidence does not indicate that clay translocation has been an important pedogenic process at this site. It is noteworthy, though, that the common horizon sequence for the Gleyed Eluviated Black Chernozemic subgroup includes a Btjgj horizon. In fact, elsewhere at the Ellerslie Research Station, analyses of pedons of both the Orthic and Gleyed Eluviated Black subgroups indicated the presence of Bt and Btg horizons with rare to occasional argillans (Crown and Greenlee, 1978; McKeague et al., 1978). Textural B horizons are typical of the Malmo soil series to which these pedons belong (Bowser et al., 1962) and may occur in pedons of the Orthic subgroup that lack eluvial horizons (Dudas and Pawluk, 1969b). Thus, the study site pedon shows less development of textural differentiation than is usually found in both the map unit and the subgroup to which it belongs. Whether, in fact, textural B horizons in Chernozemic soils are exclusively a product of translocation or partially the result of physical breakdown of silt-sized micaceous minerals is a question deserving of further research.

The parent material stratification also poses problems in evaluating the degree of mineral weathering. For the fine sand fraction in a homogeneous parent material, the sodic to calcic feldspar ratio should narrow with depth as a result of weathering since the latter mineral is less resistant.

Such a trend was not evident for this pedon, likely because of inherited variability in the stratified parent material. Although comparison of a single particle size fraction should reduce the variability, it is possible that more than one sediment source contributed to the Lake Edmonton glaciolacustrine basin.

If any mineral weathering has occurred in this pedon, it should be most evident in the clay fraction, by virtue of its larger surface area and reactivity. As noted earlier, though, between-horizon variations were subtle and showed few patterns of mineral abundance that could be attributed to weathering. The exceptions were the indications of an undetermined but minor amount of vermiculite that appeared to increase in abundance toward the surface, as well as the presence of a small quantity of an Al-hydroxy intergrade mineral in the Ah2 horizon. These results agree with those of other studies which have generally found a low degree of weathering in the clay fractions of Chernozemic soils (e.g. Kodama, 1979; Dudas and Pawluk, 1969b; Redmond and Omodt, 1967). The suggestion has been made by some workers that "illitization" of smectite occurs in the upper horizons of Chernozemic soils, as evidenced by higher K_2O contents (Dudas and Pawluk, 1969b; St. Arnaud and Mortland, 1963). While such a process might be operating in this pedon, it is possible that any evidence for it would be obscured by the lack of parent material homogeneity.

The clay-rich texture of the parent material, which results in a low hydraulic conductivity, is likely responsible for the hydromorphic features present in the pedon. Mottling from the $\overline{\text{Aegj}}$ downwards into the Ckgj horizons, together with the extractable Fe data, indicate redistribution of that element. The pattern of Fe_d concentration in the pedon shows a greater similarity to that existing in Gleysolic than Chernozemic soils (Stonehouse and St. Arnaud, 1971), that is, a greater degree of Fe removal from the A horizon. Black Chernozemic soils usually show some Fe eluviation from the Ah horizon, even in well drained Orthic pedons (e.g. Dormaar, 1978; St. Arnaud and Whiteside, 1964), but the extent of its occurrence at this site, plus the degree of mottling, suggests some hydromorphic influence. Stonehouse and St. Arnaud (1971) found a high correlation (R=0.84) between Fe_d and clay content in Chernozemic soils and interpreted this to mean that Fe had migrated with clay. However, for this pedon, the relationship between the same two variables was weaker (R=0.70), suggesting that another mechanism was responsible, such as periodic reduction and mobilization of Fe. As in the Chernozemic pedons analyzed by Stonehouse and St. Arnaud (1971), the amount of amorphous Fe oxides (Fe_o) was small and varied little with depth, providing further evidence of the limited degree of weathering. Since Fe_p indicates the amount of organic-complexed Fe, its peak abundance is, as expected, in the Ah horizon which has the highest organic matter content. Chernozem humus is not susceptible to eluviation, so the Fe_p values reflect this immobility by decreasing rapidly with depth.

The three Al extractions showed a higher degree of similarity than did those for Fe, bearing out the contention of Blume and Schwertmann (1969) that one extraction method is usually sufficient to indicate the distribution and abundance of free Al oxides in soils. The amounts present

are low, in agreement with those reported in the literature for Chernozemic soils (e.g. Stonehouse and St. Arnaud, 1971; Peters et al., 1978) and point out again the lack of weathering in these soils.

Although Mn was not studied quantitatively, its occurrence as mangans in the C horizons suggests that it has migrated further than Fe. Such behaviour is to be expected, since the degree of reduction in the environment needed to solubilize Mn (IV) compounds is much less than that required for Fe (III). For example, at pH4, Fe_2O_3 will be reduced to Fe^{2+} at an Eh of approximately +0.4 volts, while reduction of MnO_2 to Mn^{2+} occurs at an Eh of approximately +0.9 volts (Krauskopf, 1979).

5.3 Process Studies

5.3.1 The Physical Environment

The climate of the Edmonton area shares its continental characteristics with the rest of the Prairie region. The main features of this climatic region are a large temperature range between summer and winter and unevenly distributed precipitation, with little falling during the winter (Hare and Thomas, 1974). Bowser et al. (1962) provided the following climatic summary for the Edmonton area. July is the warmest month, averaging 16.4°C, while the mean summer temperature (May through September) is 13.3°C. January is the coldest month, averaging -14.4°C, while the mean winter temperature (November through March) is -8.9°C. The average frost-free period is about 100 days and the growing season lasts an average of about 175 days. Mean annual precipitation is 44.5 cm, with recorded extremes of 22.9 and 76.2 cm. During 75% of the years of record, the range was between 35.6 and 53.3 cm. On average, half of the annual precipitation falls during the June through August period. About 70% of the annual

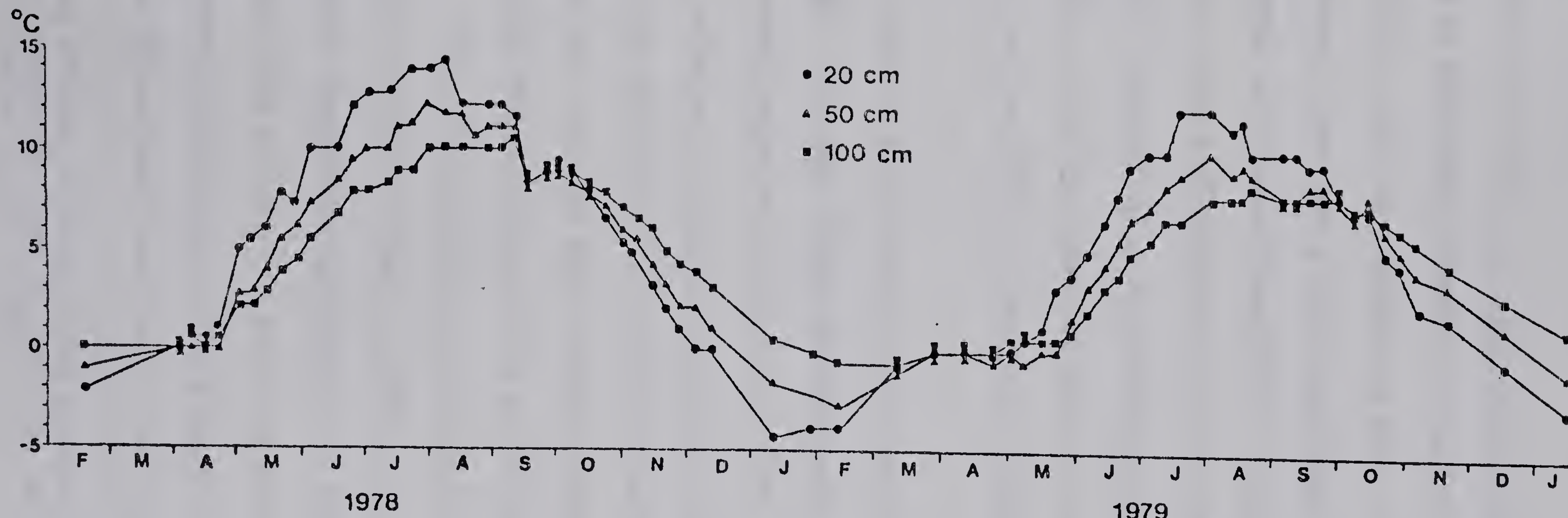

Figure 6. Soil temperatures at 20, 50, and 100 cm, February, 1978, to January, 1980.

precipitation is in the form of rain, with the rest coming as snow, averaging 127 cm.

According to the map, Soil Climates of Canada, (Clayton *et al*., 1977) Edmonton lies on the border between two subdivisions of the Cryoboreal temperature class: cold and moderately cold. The temperature requirements for the Cryoboreal class are a mean annual soil temperature (MAST) of 2-8^{o}C and a mean summer soil temperature (MSST) of 8-15^{o}C. For the 50 cm temperature readings from the study site, MAST and MSST are calculated to be 4.7^{o}C and 9.8^{o}C, respectively, for 1978 and 2.5^{o}C and 7.4^{o}C, respectively, for 1979.* Thus, these two years of data fit the Cryoboreal temperature class, except for the 1979 MSST which was slightly below the permitted minimum.

Similar rates of fall freezing of the soil occurred in both 1978 and 1979 (Figure 6). The last frost remained until the final week of April in 1978 and one month later in the spring of 1979. The cooler spring conditions in 1979 were reflected in the persistence of snow cover; as late as April 16, 10 cm of snow remained at the Ellerslie meteorological station. Spring warming of the soil occurred both from the bottom up and the top down, leaving the final frozen zone at 45-70 cm in 1978 and 50-90 cm in 1979, respectively. Thus, the period during which soil temperatures exceeded 0^{o}C at all depths lasted 171 days in 1978 and 162 days in 1979.

The greatest temperature gradients with depth occurred during the periods of early fall cooling and early spring warming. In both years, the mid-September to mid-October period showed almost isothermal conditions throughout the 183 cm depth monitored. During the 1-1½ months prior to

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*MAST = average of four readings at 50 cm equally spaced throughout the year. MSST = average of June 15, July 15, and August 15 readings at 50 cm (Soil Survey Staff, 1975).

complete thawing of the soil, temperatures stabilized at close to $0^{\circ}C$ throughout the profile.

The yearly cycle of temperature variations decreased in amplitude as depth increased (Figure 6). Peak summer temperatures occurred in late July and early August for the 20 and 50 cm depths, with the 100 cm readings showing a gentler plateau extending through August and September. Winter temperatures displayed a similar lag effect, with minimum values at 20 cm occurring about 1½ months before the minimum at 100 cm. Similar lag effects were noted by Toogood (1979) during an 8 year study of soil temperatures elsewhere at the Ellerslie Research Station.

The changes observed in soil moisture status during the April-November periods of 1978 and 1979 reflected a number of factors: depth of winter snowpack, precipitation amounts and frequency, interception by vegetation and litter, and moisture consumption for evapotranspiration. The first two factors differed considerably between the two years of records and resulted in contrasting patterns of soil moisture status.

Precipitation was higher in 1978 than in 1979 for six of the eight months during which soil moisture was monitored (Table 14). Total precipitation was 47% higher during April-November, 1978, as compared to the similar 1979 period. However, snow cover at the study site persisted until mid-April in 1979, but had disappeared by late March at the meteorological station. At the forested study site, complete snowmelt required an additional 2-3 weeks.

During both years, the H horizon always had both the highest moisture content and the widest moisture status fluctuations, reflecting the high water-holding capacity of organic matter and the greater demands of evapotranspiration in the rooting zone (Figures 7 and 8). The Ah horizon

Table 14

STUDY PERIOD PRECIPITATION,
1978-1979,
ELLERSLIE RESEARCH STATION*
(mm)

Month	1978	1979
April	19.2	32.7
May	59.8	31.4
June	38.4	78.0
July	113.8	97.2
August	93.7	29.0
September	102.7	33.6
October	14.0	11.6
November	26.7	5.0
Total	468.3	318.5

*Data provided by Division of Meteorology, Dept. of Geography, University of Alberta.

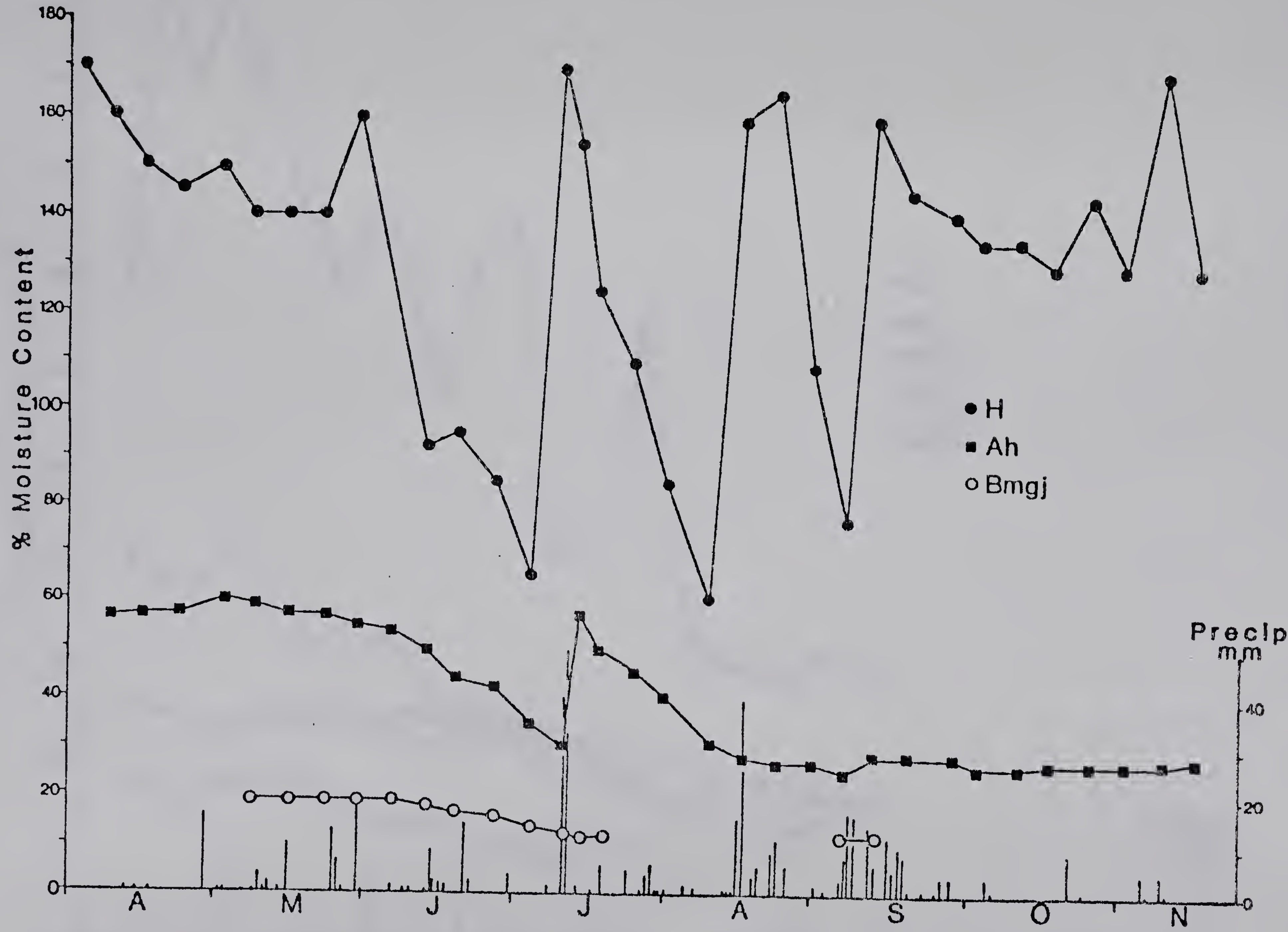

Figure 7. Soil moisture regime and precipitation, April to November, 1978.

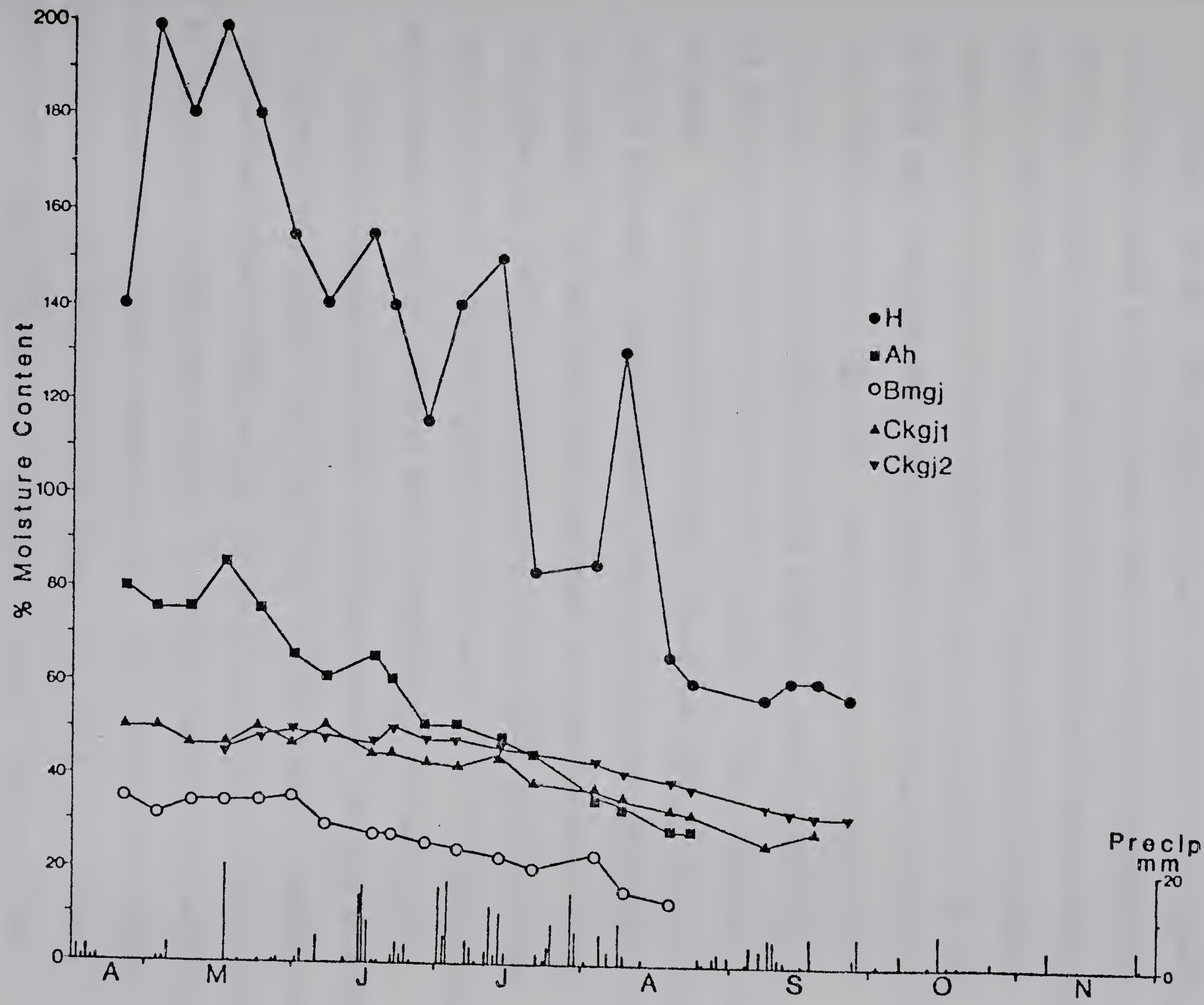

Figure 8. Soil moisture regime and precipitation, April to November, 1979.

usually showed the next highest moisture contents and amplitude of variation. The B and C horizons showed lower moisture contents, often too low to be measured by the moisture cells, and very subdued fluctuations.

In April and May, 1978, moisture levels were high in the H horizon, and were maintained by both rainfall and a lack of plant demands until late May. Rainfall was lower in June and evapotranspiration likely much greater, resulting in a sharp decrease in H horizon moisture content. During July and August, evapotranspiration would have been at its peak and only two rainstorms were large enough to restore the H horizon moisture content to its spring level: July 11-12 (9 cm) and August 16-17 (5.5 cm). September precipitation was almost as great as that in July and the H horizon moisture content was restored to 160% by the middle of the month. With the cessation of plant growth and reduction in transpiration in September, moisture content remained high until the end of the fall season. For the Ah horizon, the trend was one of gradual drying during the same period, interrupted only by the heavy rainfall of the July 11-12 storm. This event had no effect on the B or C horizons, with only sporadic readings obtained from the two deepest moisture cells.

A heavier snowpack and higher April precipitation resulted in higher spring moisture contents for all horizons in 1979. However, summer rainfall was lower than in 1978 and the individual events were smaller (Figure 8). Even after plant growth ceased, the greatly reduced fall precipitation was insufficient to raise moisture levels. Thus, all horizons showed a progressive drying trend during the summer, with the H horizon again having the widest fluctuations in moisture content. Since the lower summer precipitation of 1979 was closer to the average pattern for the Edmonton area, the soil moisture behaviour for that year was probably

more representative of the normal conditions at the site.

Interception of precipitation by litter and the tree canopy tends to reduce the supply of water entering the soil. For example, several days of light rainfall in June, 1978, had little effect on the moisture depletion trend in the H and Ah horizons, so interception losses were likely considerable. Parker (1978) found that such losses accounted for 8-9% of total rainfall for Populus tremuloides stands in northeastern Alberta, with the net rainfall being greater for larger events. Zinke (1967) reported that hardwood forest canopies can store as much as 2 mm of precipitation.

The pattern of progressive moisture depletion during the growing season is typical of forest soils (Zahner, 1967) and has important implications for soil genesis. Because of interception and transpirational demands, there were only one or two rainfall events per season that were large enough to change the moisture content of the mineral soil horizons. Therefore, the frequency of leaching events is quite restricted, perhaps limited to the early spring period during snowmelt and thawing of the soil, in addition to infrequent major rainstorms.

Piezometric head values showed gradual seasonal changes during the study (Figure 9). At all times, the head was highest in the 10.4 m piezometer, while the 4.5 m piezometer was dry throughout the period of record. This indicated upward movement of groundwater, which likely moved into a more permeable sand and gravel layer at approximately 4.5 m from the surface, to discharge elsewhere in the landscape. During 1978, head values fell between mid-June and late September and then rose more rapidly until the cessation of measurements in late November. In the spring of 1979, head values were higher than at any time during the

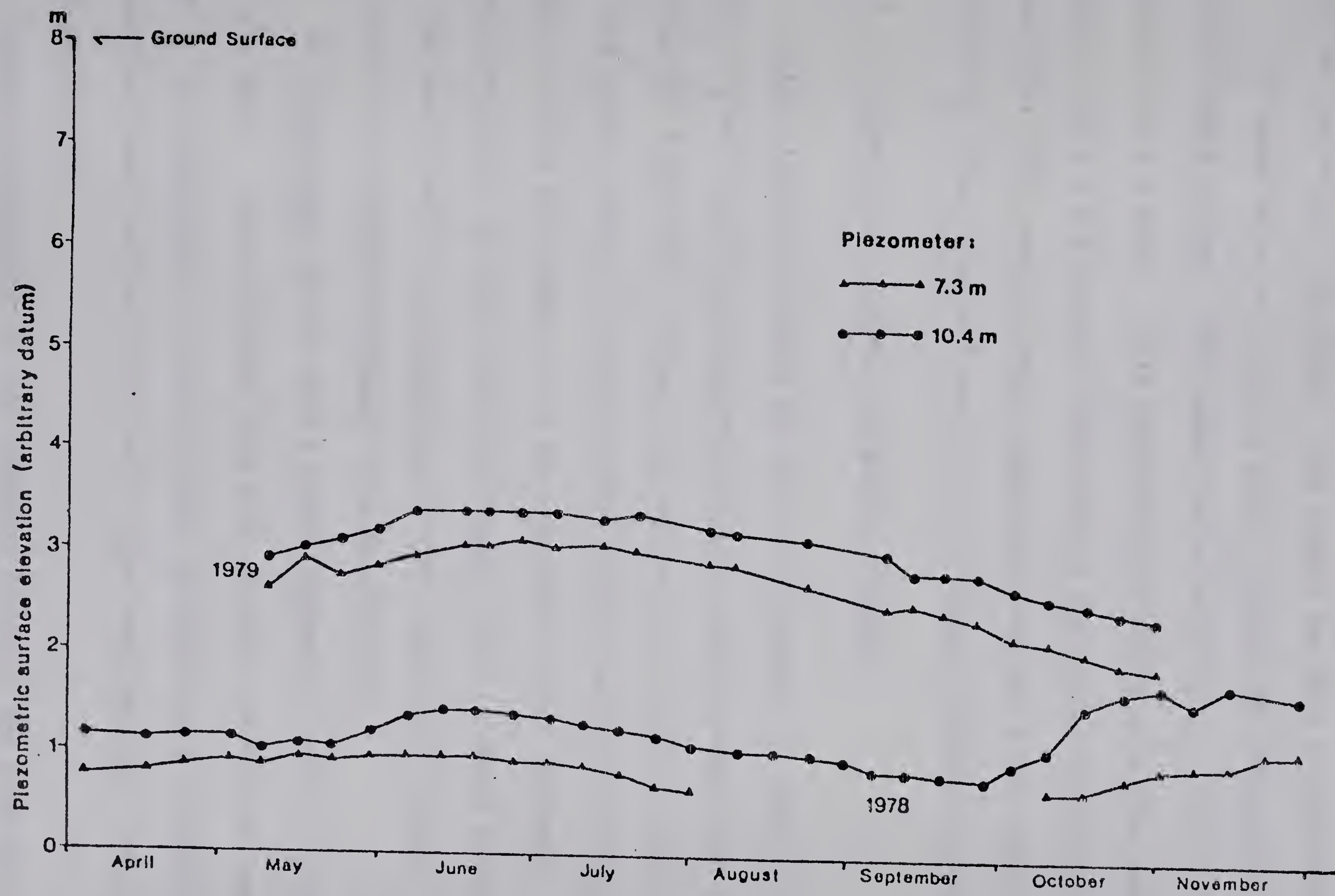

Figure 9. Piezometric head readings, 1978 and 1979.

previous year and this rising trend continued until late June. The rising piezometric head from late 1978 until mid-1979 probably was a delayed result of the higher than normal late summer and fall precipitation in 1978.

The water table well (3.6 m deep) remained dry throughout the study period except for May, 1979, when the water table was within 1-1.5 m of the surface for 2-3 weeks. This condition was likely a perched water table because the 4.5 m piezometer remained dry during the same period. Thus, except for this short episode, groundwater was not close enough to the surface to affect the soil profile, even though discharge was occurring.

5.3.2 Litterfall: Biomass and Nutrient Content

Total litterfall during the May 17 to October 31, 1979, period totalled 3537.8 kg ha^{-1}, with most occurring between mid-September and mid-October (Table 15, Figure 10). A small spring peak of litterfall consisted mostly of bud scales, seeds, and other reproductive structures of *Populus balsamifera*. During the summer, from mid-June until early September, litter consisted mostly of green leaves and small twigs released during high winds. Of the year's total, *P. balsamifera* litter accounted for about 75%, most of which fell from mid-September through mid-October.

P. tremuloides leaves were a minor component, comprising only 7.1% of the total. The peak input from this species occurred one week after that of *P. balsamifera*. Since no large specimens of *P. tremuloides* grew above the traps, most of these leaves blew into the site from the edge of the wooded area.

Cornus stolonifera was the dominant shrub at the site and accounted for 16.3% of the litter fall, with virtually all of that consisting of leaves and reproductive structures, principally fruits. This species had

Table 15 TOTAL LITTERFALL, MAY-OCTOBER, 1979

Species	Litter type	kg ha^{-1}	% of total
P. balsamifera	Leaves and reproductive structures	2483.0	70.2
	Bark and twigs	194.0	5.1
P. tremuloides	Leaves	250.0	7.1
C. stolonifera	Leaves and reproductive structures	575.0	16.3
	Twigs	9.3	0.2
Rosa sp.	Leaves	7.3	0.2
	Twigs	3.0	0.1
Prunus sp.	Leaves	8.3	0.2
Rubus sp.	Leaves	2.3	0.1
Corylus cornuta	Leaves	5.6	0.2
TOTAL		3537.8	

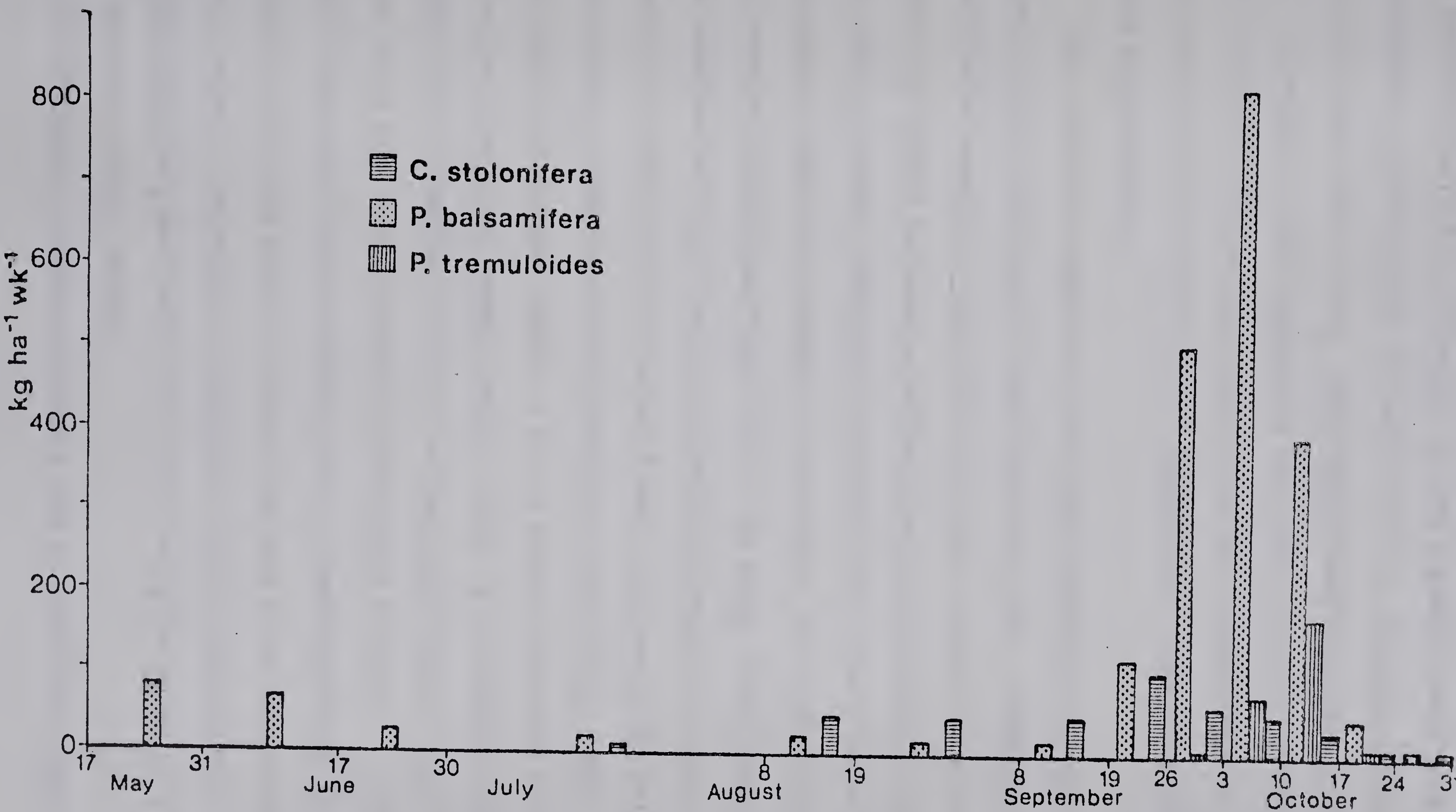

Figure 10. Weekly rates of leaf litterfall, 1979.

its maximum rate of leaf litterfall in late September, with most of its contribution occuring before the peak fall of _P_. _balsamifera_ leaves. In fact, more than half of the _C_. _stolonifera_ leaves had fallen by mid-September. Although no litter collections were made during 1978, it was also observed during that year that leaves from this species began to turn colour and fall well before those of _Populus_ spp., suggesting that such timing may be typical.

Three important components of the annual addition of organic matter to the soil were neglected by these litter trap data. Although herbaceous plants had a coverage of about 20% in the area in which litterfall was sampled, no attempt was made to estimate their contribution to the total. Root death is another unestimated component of the organic matter budget, but because of practical problems of measurement, few studies exist of root turnover in deciduous forest soils. Cox _et al_. (1978) found that in a Tennessee deciduous forest, root death and decay (< 0.5 cm diameter) accounted for about 2/3 of the annual biomass returned to the soil. In both deciduous forests and pine plantations, Harris _et al_. (1977) found that annual root turnover (mostly < 0.5 cm diameter) amounted to about 25% of the below-ground organic matter. Thus, it is probably a reasonable estimate that in deciduous forests, the root contribution to soil organic matter is at least equal to that of above-ground litterfall. A final point is that since litter collection was not continued during the winter and early spring, the amounts of branch and twig litter were considerably underestimated. For example, Cragg _et al_. (1977) found that most of the _P_. _tremuloides_ twig and branch litter fell during the winter at a site in the Kananaskis valley, Alberta.

The litterfall estimates from this site are within the range reported

for other aspen forests in the cool temperate zone. In southwestern Quebec, Coldwell and DeLong (1950) reported annual rates of 1123 to 2291 kg ha^{-1} of leaf litter. Cragg et al. (1977) reported a biomass of 1734 kg ha^{-1} yr^{-1} in the form of leaf litter from a 69 year old aspen stand in the Kananaskis valley. In the U.S.S.R., Rodin and Bazilevich (1967) measured leaf litterfall rates ranging from 3620 to 4470 kg ha^{-1} yr^{-1} in aspen stands ranging from 10 to 50 years old.

Since leaves and reproductive structures of P. tremuloides, P. balsamifera, and C. stolonifera accounted for most of the litterfall, only these materials were selected for study of nutrient return. Ca, Mg, and K contents were examined because these elements accounted for virtually all of the exchangeable cations in the soil. Seasonal changes in leaf nutrient content are averaged out in these data since the subsamples for chemical analysis were composites weighted according to the litterfall in each sampling period for that particular species. Such changes can be significant; James and Smith (1978) found that for P. tremuloides leaves, the Mg content increased twofold and the K content fluctuated by a similar factor during the growing season.

Table 16 indicates that both Populus species were similar in Ca and Mg contents, but that the K content of P. tremuloides was closer to that of C. stolonifera. The latter had the lowest K and Mg contents but was about 30% higher in Ca content than the average of the two Populus species. In terms of total nutrients returned, P. balsamifera was dominant because of its large biomass. However, between-species differences in nutrient content can obviously alter the total amount returned by a species. For example, while C. stolonifera leaves comprised only 17.4% of the leaf litter contributed by the three dominant species, this source returned

Table 16. NUTRIENT CONTENTS OF LEAF LITTER

Species	Ca			Mg			K		
	%	kg ha^{-1} yr^{-1}	% of total*	%	kg ha^{-1} yr^{-1}	% of total*	%	kg ha^{-1} yr^{-1}	% of total*
P. balsamifera	2.17	52.6	71.9	0.36	8.76	77.9	0.73	17.80	80.3
P. tremuloides	2.08	5.1	7.0	0.36	0.88	7.8	0.60	1.47	6.6
C. stolonifera	2.77	15.5	21.2	0.29	1.60	14.2	0.52	2.90	13.1
Total		73.2			11.24			22.17	

* % of total returned by leaf litter of three dominant species.

21.2% of the Ca.

The important contribution of _C. stolonifera_ to Ca cycling at this site appears similar to the findings of Thomas (1969). A related eastern U.S. species, _C. florida_, comprised only 0.2% of the aboveground biomass in the pine plantation studied, yet contained 1.8% of the total aboveground Ca in the stand.

The nutrient quantities returned in litterfall at the study site fit into the higher range of values reported for aspen forests (Table 17). For the Alaska and Kananaskis aspen stands, site descriptions provided by the authors indicated that the shrub stratum was much less important than at Ellerslie; this may account for part of the difference in nutrient return.

5.3.3 Litter Decomposition

Decomposition rates of the three dominant leaf litter types were studied by two methods: incubation, with measurement of CO_2 production in the laboratory, and by the litter bag method at the study site.

The incubation experiment showed the _C. stolonifera_ leaves to decompose approximately twice as fast as the _Populus_ spp. leaves (Figure 11). The latter showed similar rates, with 25.7 and 21.4% of the carbon content of _P. balsamifera_ and _P. tremuloides_ leaves, respectively, having been realeased as CO_2, compared to 42.4% for _C. stolonifera_ during the same 72 day period. Since all leaf materials were ground to the same size and were incubated under uniform temperature and moisture conditions in the absence of soil mesofauna, these results should indicate their relative palatability as microbial substrates. Accordingly, _C. stolonifera_ leaves seem to contain more readily metabolized substances than do the

Table 17

COMPARATIVE DATA FOR NUTRIENT RETURN
IN ASPEN FOREST (kg ha^{-1} yr^{-1})

Location and Particulars	Ca	Mg	K	References
U.S.S.R. (total litter)	66.0 - 105.1	9.9 - 14.2	22.4 - 84.7	Remezov and Pogrebnyak (1969)
Alaska (50 years old, leaf litter only)	48.0	6.8	9.6	Van Cleve and Noonan (1975)
Alaska (120 years old, leaf litter only)	40.2	6.2	6.2	Van Cleve and Noonan (1975)
Kananaskis, Alberta (leaf litter only)	35.8	5.7	10.6	Cragg _et al_. (1977)

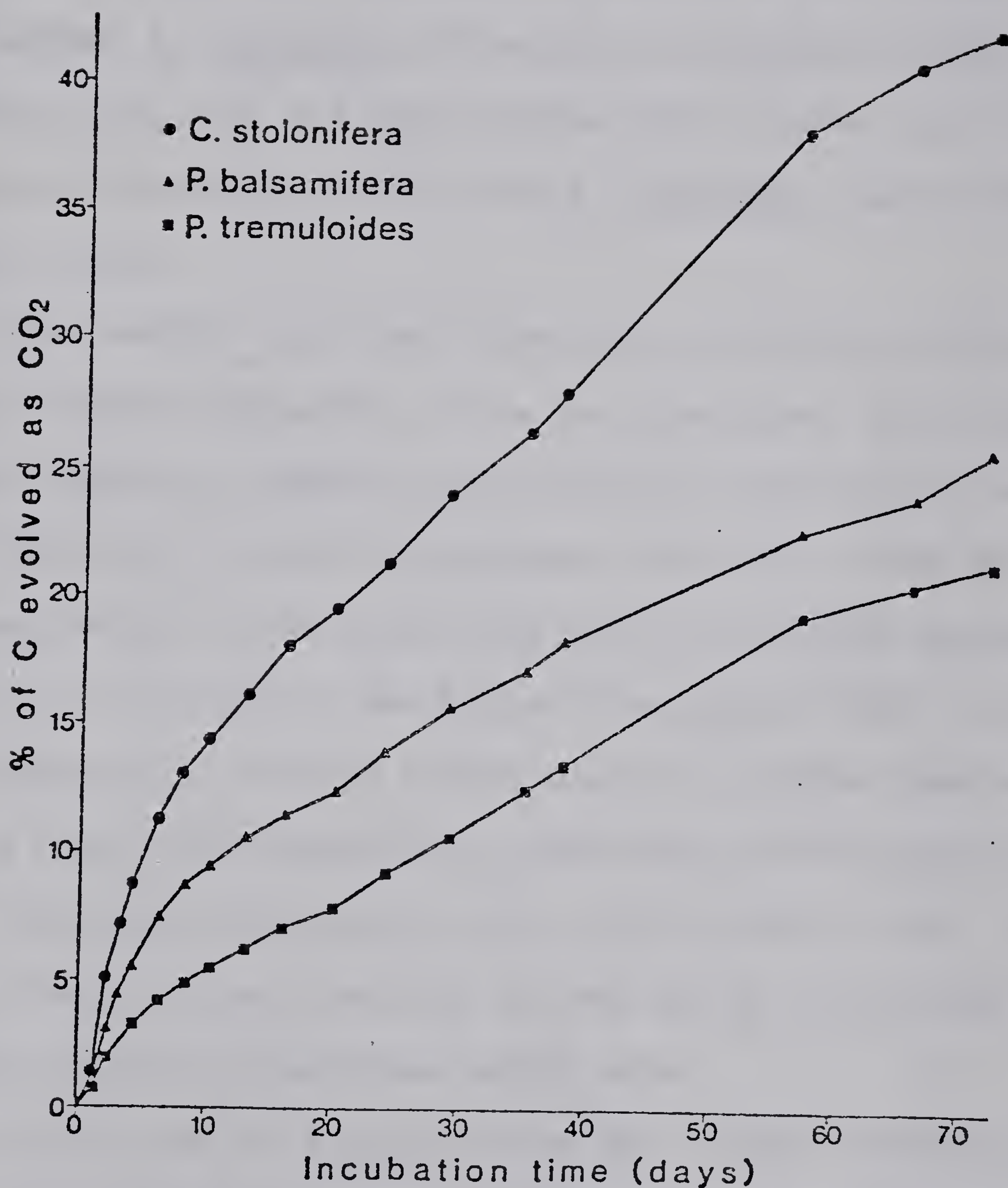

Figure 11. Incubation experiment: rates of CO_2 production from leaf litter materials.

Populus spp. leaves.

Three noteworthy points emerged from the litter bag experiment (Figure 12). (1) All three species lost a significant proportion of their initial weights over the winter months. (2) Although the differences were not statistically significant because of the limited number of samples, C. stolonifera litter bags exhibited the highest rate of weight loss. (3) All three species showed similar rates of weight loss until mid-summer, after which C. stolonifera leaves seemed to break down faster.

Over-winter weight losses from litter bags are usually attributed to leaching of soluble components, since low temperatures limit faunal and microbial activity. Anderson (1973) found that leaching can account for most of the weight losses from deciduous tree leaves during the first year of decomposition. Considerable over-winter losses from deciduous leaf litter were also found in New England (Gosz et al., 1973). However, unlike the latter site, at which a thick snowpack prevented freezing of the soil and litter, the Ellerslie site experienced frozen conditions in the litter layer from mid-November, 1978, until mid-March, 1979. Therefore, it is likely that most leaching occurred during the late March through April period of snowmelt and spring rains.

Weight loss by the two Populus species was a linear function of time, with simple regression giving a high correlation coefficient for both species ($R = 0.93, 0.97$) (Figure 12). Although not different at a statistically significant level, the rates of weight loss for leaves of these two species were both slower than that observed for C. stolonifera. P. balsamifera leaves showed the slowest rate, probably reflecting their size which is approximately four times larger than leaves of

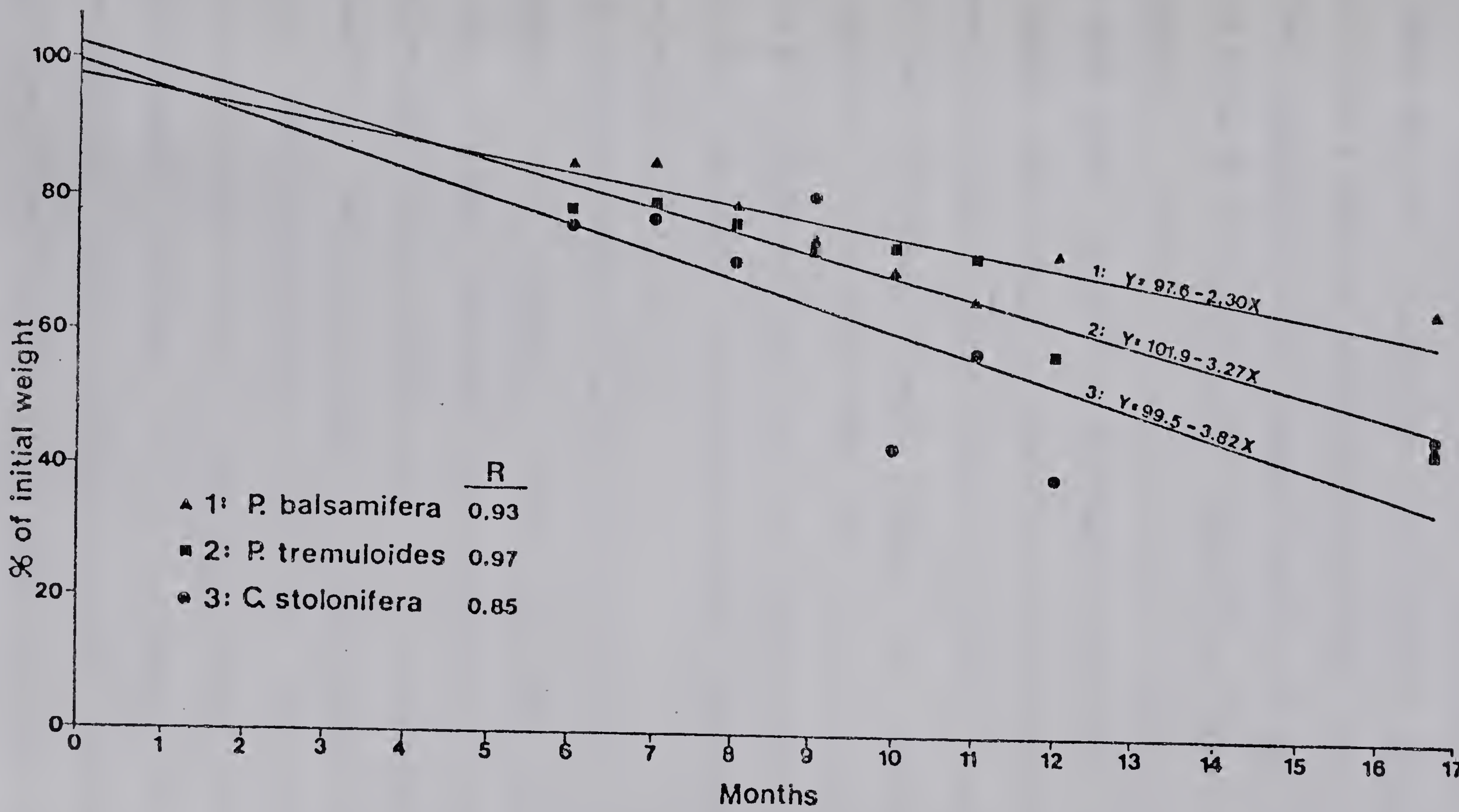

Figure 12. Litter bag weight loss rates.

P. tremuloides. P. balsamifera leaves were also thicker and less brittle than those of the other two species, characteristics that would slow their rate of fragmentation. By the end of the experiment, the Populus spp. leaves were still recognizable, although the P. tremuloides leaves showed a greater degree of skeletonization (Plates 10 and 11).

The C. stolonifera litter showed the most rapid rate of weight loss, with the slope of its regression line being 65% steeper than that of P. balsamifera. However, the R value is lower, indicating that the linear model gave a poorer fit. This particularly seemed to be the case after August, when there was a sharp drop in the recovered weight and an increase in the month-to-month variation in percentage weight remaining. At that time, the effects of faunal ingestion and fragmentation became quite pronounced. By November 1979, little remained other than leaf petioles and midribs, with the rest of the tissue reduced to small fecal pellets which were easily lost through the mesh, thus increasing the apparent weight loss (Plate 12).

The regression lines for weight loss indicated 37.4% and 30.0% for P. tremuloides and P. balsamifera, respectively, after twelve months. These rates are higher than those found by Lousier and Parkinson (1976) in the Kananaskis valley, Alberta, where the corresponding values were 26.2% and 21.2%. The cooler climate and shorter growing season of the latter site likely accounted for the slower rate. The authors found that the faster weight loss by P. tremuloides corresponded to a higher content of leachable substances. The rates obtained at the Ellerslie site are closer to those found by MacLean and Wein (1978) for P. tremuloides leaves in New Brunswick: 38.3% weight loss after one year. Comparable data could not be found for C. stolonifera, but it is of

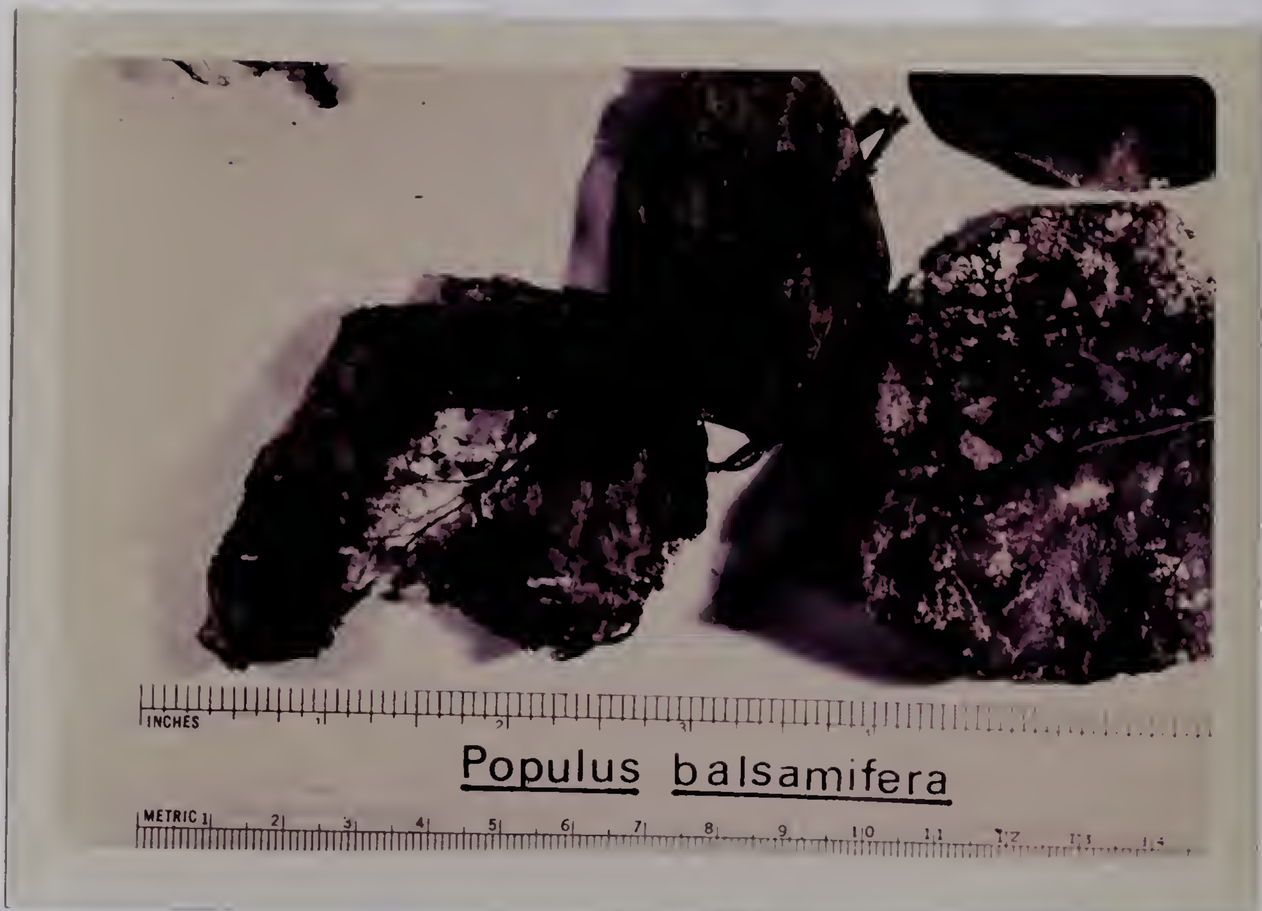

Plate 10. Populus balsamifera litterbag material, recovered after 13 months.

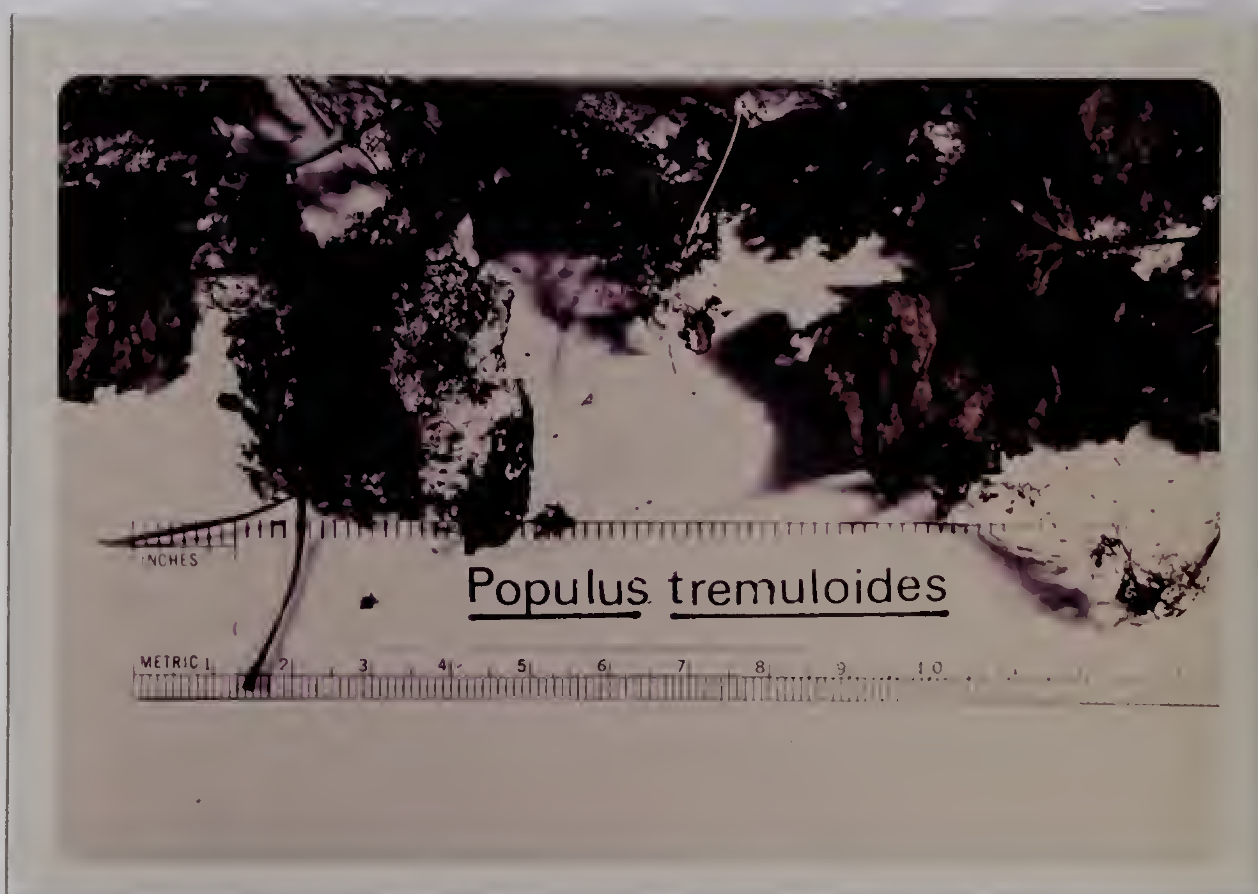

Plate 11. Populus tremuloides litterbag material, recovered after 13 months.

Plate 12. Cornus stolonifera litterbag material, recovered after 13 months.

interest that for the related C. florida, leaf litter was found to have a half-life of only 0.6 years (Ausmus and Witkamp, 1974) which was less than any other value cited by Lousier and Parkinson (1976).

It was obvious that faunal activity was a major reason for the rapid physical breakdown of C. stolonifera leaves. In addition, certain differences also existed in the microbial populations colonizing these three litter types (Table 18). Pigmented (orange or pink) colonies comprised half or more of the bacterial colonies counted on the Populus spp. plates, but only about 25% of those on the C. stolonifera plates. Such pigmentation is characteristic of the genus Cytophagas which has a spreading habit of growth adapted to the exploitation of extensive, lower quality food resources (Dr. F.D. Cook, Dept. of Soil Science, University of Alberta, personal communication). This suggests that the C. stolonifera leaf litter is a higher quality substrate for bacteria, in addition to being more palatable to mesofauna.

Comparison of nutrient contents in the litterbag samples during the first year of the experiment revealed different trends both according to the species and the element studied. Calcium contents of litterbag samples behaved differently in C. stolonifera, which showed wide fluctuations converging on the original value, compared to Populus spp., which displayed an increasing Ca content after several months (Figure 13). Magnesium contents showed less fluctuation in the C. stolonifera than Populus spp. samples. In the latter, fluctuations for both species were similar in direction and magnitude over the duration of the experiment. Overall, the Mg content remained close to the original value for C. stolonifera and about 50% lower for Populus spp. Potassium showed both the smallest difference in its dynamics between the three litter types, and the most

Table 18. MICROBIAL POPULATIONS IN LITTER BAG MATERIALS (OCTOBER, 1979)[1]

Leaf type	Bacteria A[2]	Bacteria B[3]	Fungi
P. tremuloides	1.1×10^8	8.3×10^7	1.3×10^7
P. balsamifera	2.4×10^8	2.4×10^8	2.0×10^6
C. stolonifera	7.2×10^7	2.2×10^8	2.9×10^7

[1] populations expressed as numbers per g, on a moisture-free basis

[2] A - formed coloured colonies (orange, pink)

[3] B - formed white colonies

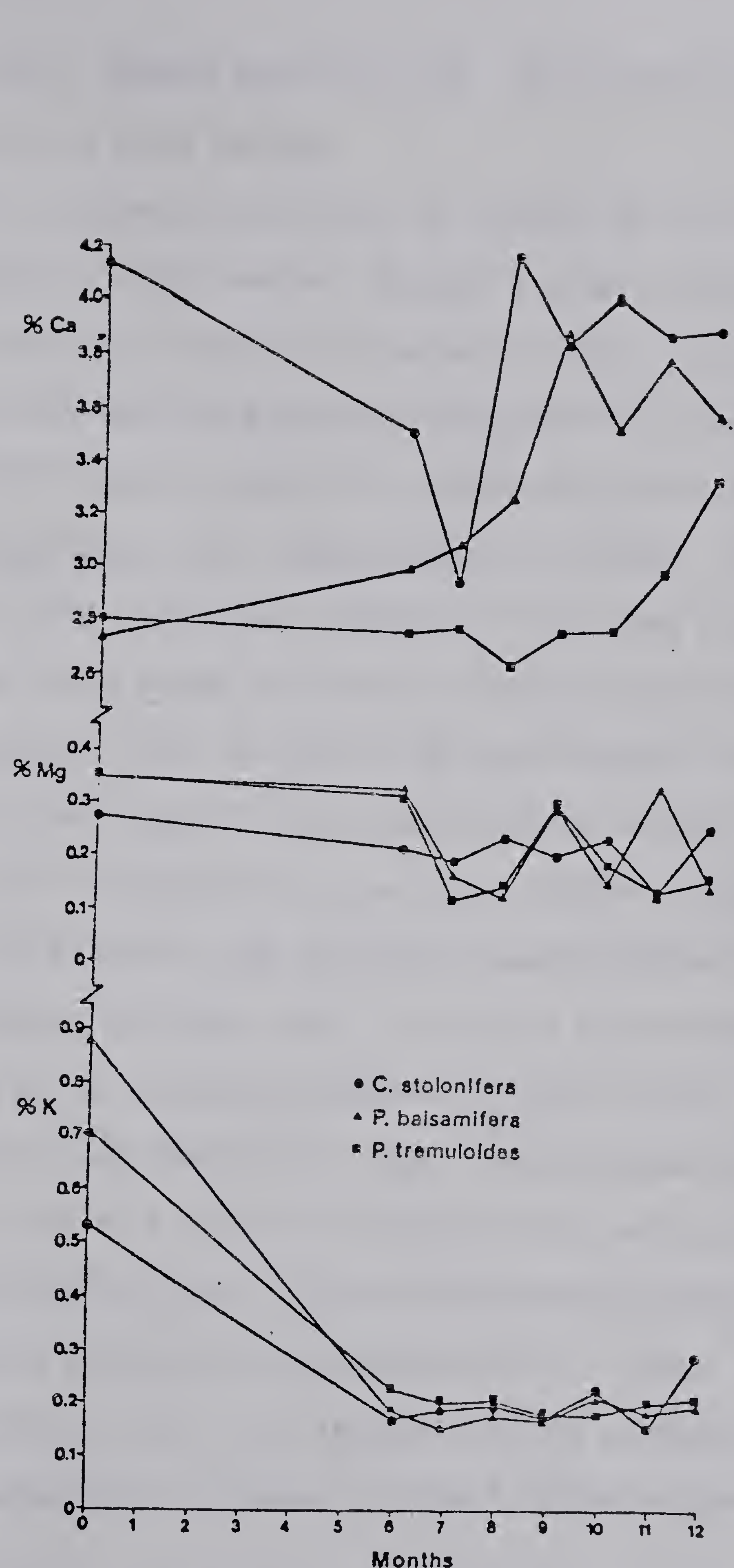

Figure 13. Nutrient content of decomposing leaf litter: Ca, Mg, and K.

rapid rates of loss, ranging from 45 to 75%. Most of this loss occurred during the winter and early spring.

These results generally correspond to patterns of litter nutrient release documented by other workers. MacLean and Wein (1978) found that the Ca content generally increased in hardwood litter, including that of P. tremuloides, after between eight and twelve months of decomposition in litterbags. Similar results were also obtained by Lousier and Parkinson (1978) for P. tremuloides and P. balsamifera leaf litter. Gosz et al. (1973), Attiwill (1968) and Thomas (1970) all found that Ca was the least mobile element of these three, with overall losses closely related to the rate of tissue weight loss. Ca tends to be concentrated in the structural components of the leaves and is less susceptible to leaching. Literature data indicate that K is rapidly lost and shows smaller between-species differences in its behaviour than do other elements (Attiwill, 1968; Gosz et al., 1973; MacLean and Wein, 1978). This loss is attributed to leaching because K is not a structural component of plant tissue. Although Gosz et al. (1973) found that Mg loss rates from deciduous leaf litter were similar to those of K, MacLean and Wein (1978) and Lousier and Parkinson (1978) observed rates that were intermediate between those of K and Ca, a finding similar to that obtained in this study. Therefore, the mobility ranking, K > Mg > Ca, indicated by the present study is typical of the pattern for deciduous tree leaf litter in general.

5.3.4 Water Chemistry: Precipitation, Throughfall, Stemflow, and Lysimeter Leachates

Distinctive patterns of variation were displayed by the selected soluble constituents in water sampled at different levels at the study site: precipitation (prior to interception by vegetation), throughfall,

stemflow, and soil leachates. Before discussing these results, it is essential to realize the constraints imposed on their interpretation by the nature of the system and the sampling problems involved.

Since the objective was to observe general trends, rather than to establish accurate chemical budgets, sampling intensity was low. Because both throughfall and stemflow have considerable spatial variation in both quantity and chemistry in a forest stand, very large numbers of collectors are needed to give a valid sample if a budget type of study is to be conducted (Parker, 1978). Chemical variability also results from the extent of dilution effects which are governed by the amount and frequency of rainfall involved in leaching substances from leaf surfaces; this matter will be discussed later. To allow for all of these sources of variability would have made for a major study in itself. Therefore, the data presented for throughfall and stemflow are intended only as general indicators of water composition at the site.

The nature of the lysimeter design and placement justifies a similar degree of caution in interpretation. Large voids, such as root channels, greatly influence rates of water movement through the soil, but such features are unpredictable in their distribution. This type of porosity would obviously affect the extent of interaction between water and soil components and therefore the characteristics of the leachate chemistry. Secondly, the walls of the lysimeter columns exclude plant roots, affecting the soil solution in at least three ways: (1) prevention of water uptake for transpiration, thus increasing leachate volumes, (2) prevention of nutrient uptake, possibly resulting in higher solute concentrations (cf. Feller, 1977), and (3) elimination of rhizosphere effects, such as root exudation and respiration. The weekly sampling schedule meant that

after heavy rainfalls, the lysimeter collection vessels would likely overflow, resulting in loss of leachate and preventing any quantitative estimate of total nutrient flux.

For the foregoing reasons, the data are presented as simply as possible -- as unweighted means for each property determined for samples from a particular source. Despite all of these qualifications, it will be seen that several trends emerged which are consistent with soil properties and our understanding of solute dynamics in ecosystems.

The mean pH was lowest in the incoming precipitation (6.33) and increased as water moved downward through the system (Table 19). The largest pH change occurred when the precipitation interacted with the tree canopy, with the result that pH values for the lysimeter leachates were close to those recorded for the throughfall and stemflow.

The major cations (K^+, Ca^{++}, Mg^{++}, Na^+) were all present at low levels in the precipitation but differed in their behaviour as water passed through the system. Substantial amounts of K^+ were added to throughfall, with lesser amounts of Ca^{++}, Mg^{++}, and Na^+, in decreasing order of abundance. Stemflow, compared to throughfall, was enriched in three of these elements, while maintaining similar concentrations of Na^+. The most dramatic increase occurred for K^+, which was approximately ten times as concentrated in stemflow as in throughfall. Ca^{++} and Mg^{++} were enriched by smaller factors in stemflow. Stemflow cation concentrations for all elements except Ca^{++} were equal to or greater than those found in the soil leachates. However, the vegetation cover above the lysimeter installation was similar to that above throughfall collector #2 and no tree trunks were close enough to contribute stemflow to the water entering the columns. Therefore, the data reported for that collector

Table 19

PRECIPITATION, THROUGHFALL, STEMFLOW, AND SOIL LEACHATE CHEMISTRY[1]

Source	pH	m.e. l^{-1}						mg l^{-1}				
		K^+	Ca^{++}	Na^+	Mg^{++}	$SO_4^=$	HCO_3^-	Si	Soluble C	Uronides	Saccharides	Phenols
P	6.33(0.45)	0.01(0.008)	0.05(0.06)	0.01(0.009)	0.01(0.007)	0.04(0.006)	0.21(0.08)	n.d.[2]	n.d.	n.d.	n.d.	n.d.
T_1	6.90(0.42)	0.36(0.25)	0.26(0.18)	0.06(0.06)	0.11(0.07)	n.d.	0.47(0.30)	n.d.	40(19.7)	12(0)	24(23.8)	20(18.3)
T_2	6.89(0.37)	0.18(0.12)	0.18(0.12)	0.04(0.04)	0.07(0.04)	0.06(0.0)	0.30(0.16)	n.d.	28(15.1)	9(2.1)	9(5.6)	8(4.6)
T_3	7.20(0.22)	0.46(0.36)	0.25(0.20)	0.06(0.06)	0.11(0.08)	0.21(0.17)	0.74(0.57)	n.d.	35(17.3)	12(3.2)	12(9.6)	14(13.5)
S_1	7.44(0.31)	3.48(1.56)	1.02(0.65)	0.04(0.03)	0.54(0.19)	n.d.	5.61(2.83)	n.d.	83(56.3)	35(16.2)	33(21.1)	21(10.0)
S_2	7.20(0.08)	2.47(1.18)	0.79(0.67)	0.07(0.06)	0.47(0.29)	0.16(0.12)	1.67(1.34)	n.d.	53(24.6)	21(5.0)	32(24.1)	31(29.5)
L-F	7.33(0.24)	0.94(0.22)	1.93(0.96)	0.08(0.04)	0.48(0.31)	0.06(0.06)	2.20(0.33)	n.d.	46(17.7)	15(6.3)	10(7.3)	4(1.4)
H	7.21(0.28)	0.49(0.21)	1.61(0.59)	0.08(0.04)	0.34(0.11)	0.28(0.11)	1.64(0.86)	n.d.	51(21.4)	13(3.4)	6(3.0)	4(2.7)
Ah	7.33(0.31)	0.26(0.15)	2.01(0.71)	0.10(0.06)	0.61(0.30)	0.40(0.12)	1.64(0.77)	6.29(4.04)	41(22.9)	12(13.3)	10(4.8)	2(1.1)

1 mean values, with standard deviation in ()

2 n.d.- not determined

are probably the best guide to the probable concentrations of solutes entering the lysimeters; generally, the lowest average concentrations in throughfall were found in #2.

Once throughfall entered the soil, cation behaviour differed according to the element. K^+, which is more abundant in the L-F leachate, decreased by about 50% between each level. This trend was not merely an artifact of the averaging of values, but was the case for virtually all individual sampling periods. Na^+ remained the least abundant cation, increasing slightly in the Ah horizon leachate. Ca^{++} and Mg^{++} behaved almost identically, each with concentrations in the L-F leachate being much greater than in the throughfall. These two cations then decreased in the H horizon leachate but reached a maximum in the Ah leachate.

Because of sample volume limitations, only two anions were determined: HCO_3^- and $SO_4^=$. Other anions were likely present: NO_3^-, which was identified by infrared analysis of concentrated samples, and Cl^-, which occurs in small amounts, averaging 0.5 mg l^{-1}, in precipitation in central Alberta (Summers and Hitchon, 1973). Bicarbonate was the dominant anion at all levels in the system and its behaviour followed the general trend of increasing cation concentration in the order: precipitation, throughfall, and soil leachate. Stemflow contained bicarbonate concentrations equal to or greater than those in the soil leachates. Sulphate concentration was lowest in the precipitation samples, 1.7 mg l^{-1}, typical of the region, (Summers and Hitchon, 1973), higher in the throughfall and stemflow, and increased progressively with depth in the soil leachates.

Silicon was determined on all Ah horizon leachate samples and on a selection of samples from other sources; only the former contained any detectable amounts. The mean concentration was 6.29 mg l^{-1} and ranged

between 0.4 and 10.35 mg l^{-1}. Because of the filtration procedure used in handling the leachates, it is possible that some undetermined part of the Si was present in colloidal forms that were not removed. Ah horizon leachates were also tested for Fe and Al, but any amounts present were below the detection limits of atomic absorption spectrophotometry.

The organic constituents in solution showed two patterns of variation. Soil leachate concentrations were either roughly equal to those in throughfall (soluble C, uronides, and saccarides) or lower, as in the case of total phenols. For all components determined, the highest concentrations were found in stemflow.

Freeze-dried leachates, throughfall, and stemflow samples collected on August 4, 1979, were examined by infrared spectroscopy. As the other analytical data indicated, these water samples possessed a complex mixture of organic and inorganic solutes. Consequently, the infrared spectra are difficult to interpret except for assessing general trends (Dormaar, 1970), in this case, the qualitative differences in the organic and inorganic solutes. The criteria used were those set out by Dormaar (1970, 1978) who interpreted similar spectra of artificial canopy drip and Ah horizon soil solutions.

As in the case of Dormaar's (1978) Ah horizon soil leachates inorganic salts (carbonates, sulphates, and nitrates) dominated the five spectra, resulting in most absorption bands for organic components being obscured (Figure 14). The 2900 cm^{-1} band for aliphatic CH_x stretch was clearly present in throughfall, stemflow and L-F leachate samples, but was not discernible in the H and Ah horizon leachates. Carbonyl absorption (1720 cm^{-1}) was not detected in any spectrum. The 1380 cm^{-1} absorption band attributed to the carboxylate anion partially coincided with the strong

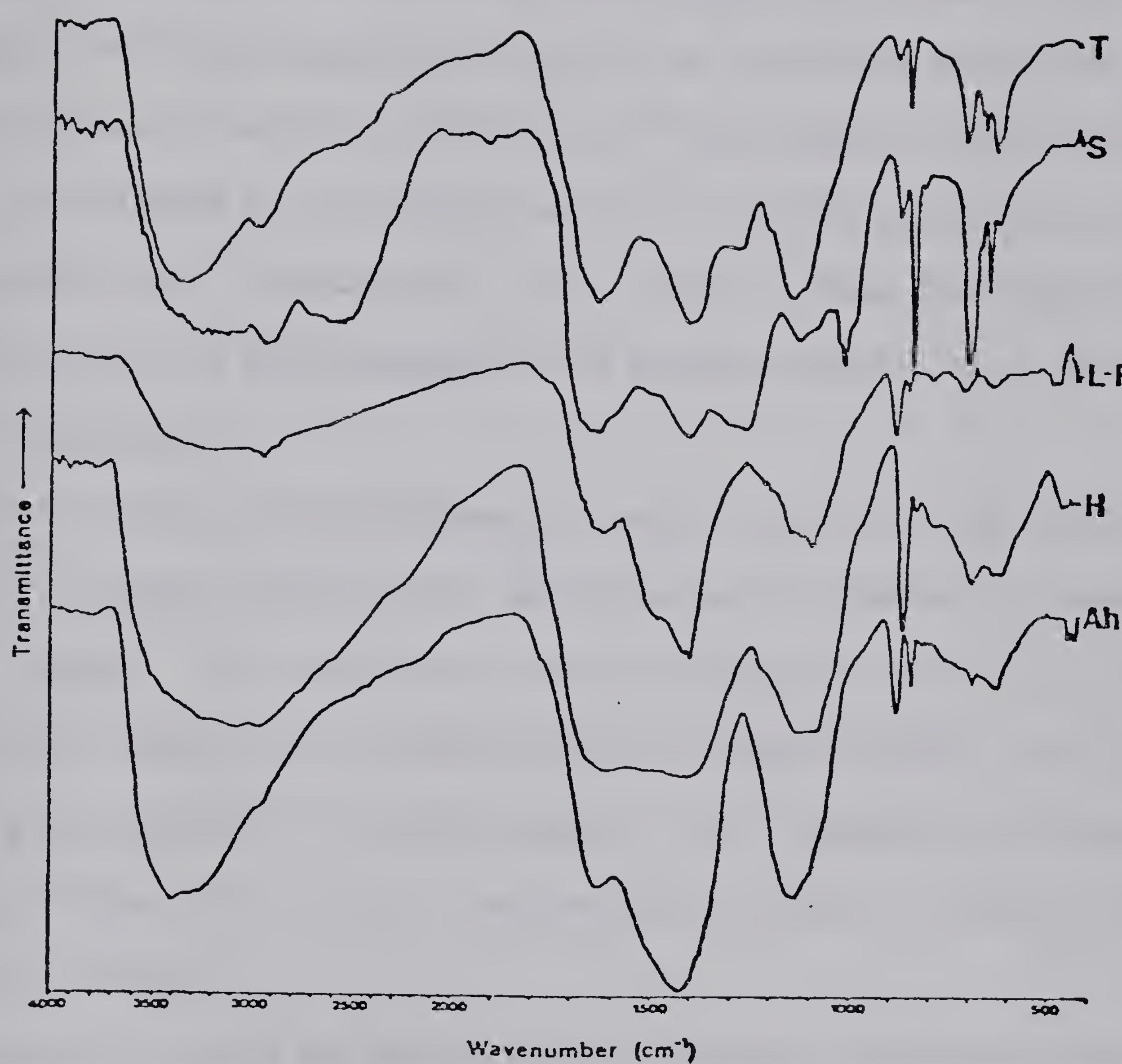

Figure 14. Infrared spectra of freeze-dried throughfall (T), stemflow (S), and lysimeter leachates.

bands from $CO_3^=$ and NO_3^- and was clearly evident only in the L-F leachate. It is noteworthy that this was consistent with the finding that the highest concentrations of uronic acids were detected in the L-F horizon leachates. Bands near 1075 and 1045 cm^{-1} were clearly evident only in the L-F leachate, but may be obscured in the other spectra by a strong $SO_4^=$ band. In a study of artificial leaf leachates, Dormaar (1970) attributed the former bands to silicate clay structures present as dust on leaf surfaces. Benzenoid (900-700 cm^{-1}) and phenol (1330-1110 cm^{-1}) bands are difficult to distinguish because of strong nearby absorptions by inorganic salts. Nevertheless, absorption in these two regions appeared to be greater in the throughfall and stemflow spectra than in those of the soil leachates.

In discussing the significance of these results for the functioning of the soil system, the data will be discussed in relation to several general issues: (1) the role of the tree canopy and litter layer in modifying the chemistry and distribution of precipitation, (2) the sources and functions of soluble organics, (3) patterns of elemental mobility in the soil, and (4) weathering processes in relation to Si, Fe, and Al release.

It was clear from the data presented that the tree canopy exerted a significant and spatially variable influence on the chemistry of intercepted precipitation. Although this phenomenon has been known for many years (Tukey, 1971), the quantitative aspects of this process and its ecological significance have only recently been studied. Virtually all organic and inorganic substances found in plants can be leached from their surfaces, particularly foliage. K, Ca, Mg, and Mn are the major inorganic nutrients leached from plants, but organic substances, notably

carbohydrates, comprise the largest proportion of the leached material.

For inorganic nutrients, a variety of studies have examined the quantities involved. The leaching losses of a given element are more closely related to where it occurs in the plant tissues than its simple initial abundance, as shown by Eaton et al., 1973. The same study found that leachability of particular elements from a hardwood forest canopy followed the order Na>S>K>Mg>Ca>N>P. While K and Na tend to occur in ionic forms and are readily leached, Ca and N are found in structural components and require leaf senescence before major losses can occur. At the ecosystem level, these differences result in contrasting patterns of behaviour for different elements. For example, Duvigneaud and Denayer-De Smet (1970) found that in a Belgian oak forest, throughfall and stemflow accounted for 53.3% of the annual K return, compared to 5.6% of the Ca return. The corresponding figures for an aspen stand in northeastern Alberta were 29% (K) and 9% (Ca) (Parker, 1978).

Mechanisms proposed for this leaching process involve cation exchange reactions and ion diffusion occurring at the leaf surface. For Ca^{++} leaching, Mecklenburg et al. (1966) suggested that (1) H^+ in the leaching solution could exchange for Ca^{++} on exchange sites in the cuticle and cell wall, and/or (2) Ca^{++} in the translocation stream could diffuse from the leaf into the leaching solution. These authors thought that this model would also be applicable to other cations (e.g. K^+). The same study also offered an explanation for the frequently alkaline reaction of the plant leachates. Carbonic acid from dissolved CO_2 dissociates, with the released H^+ displacing exchangeable cations from the leaf surface, changing the leaching water to an alkaline bicarbonate solution.

Stemflow has received somewhat less quantitative study, partly

because only a small proportion of the incident precipitation in involved: 3.9-6.1% for three eastern Canadian hardwood species (Mahendruppa, 1974). However, the concentrations of dissolved inorganics are often higher than in throughfall, principally because the duration of contact between the leaching water and the plant surface is longer (e.g. Rolfe et al., 1978). Epiphytic lichens, which occurred at varying densities on tree trunks at the Ellerslie site, are a further source of leachable inorganics, particularly of Ca^{++}, Mg^{++}, and H^{+} (Lang et al., 1976).

The throughfall and stemflow data presented earlier from the Ellerslie site appear consistent with the observations made in other studies. K^{+} was the most abundant cation in throughfall (0.18-0.46 m.e. l^{-1}) in accordance with its leachability, while Ca^{++} had somewhat lower concentrations. Since substantial over-winter losses of K^{+} were noted in the litterbag experiment, it is likely that much of the K^{+} in the soil leachates during the summer came from the throughfall, since the readily leached supplies of K^{+} had already been lost from the previous year's litter layer. Except for Na^{+}, the higher cation concentrations noted in the stemflow were to be expected as a result of the surface area and lichen effects. The ranking of abundances in the order $K^{+}>Ca^{++}>Mg^{++}>Na^{+}$ is almost the same as for throughfall, except that in the latter Mg^{++} and Na^{+} have approximately equal concentrations. These results suggested that the same controls of elemental mobility operate for both stemflow and throughfall. Similar rankings were observed by Gersper and Holowaychuk (1971) for eastern hardwoods. In agreement with the ion exchange model discussed earlier, the pH of the water rises as the precipitation interacts with the canopy; bicarbonate concentrations also rise.

As mentioned earlier, a variety of organic substances is removed by

leaching of living plant surfaces. In the study by Eaton _et al_., (1973) of an eastern hardwood forest, average concentrations of organic matter in the throughfall were of the same order of magnitude as in the Ellerslie data. Carbohydrates are a major component of throughfall organic matter, comprising up to 78%, with an average of about 25%, of the total organic matter contribution from the leaching of an oak forest canopy studied by Carlisle _et al_. (1966). For hardwood stemflow studied in eastern Canada, Mahendruppa (1974) found lower soluble organic concentrations than were the case for the Ellerslie data, with carbohydrates comprising up to 14.9% of the organic fraction. Although a precise tracing of sources would be a project in itself, it is significant that carbohydrate concentrations in the throughfall and stemflow exceeded those in the soil leachates. Therefore, this component of the soil solution may partially originate outside the soil.

Apart from their role as a substrate for microbial growth, carbohydrates in the soil have received much attention because of their role in soil aggregation. Numerous studies (e.g. Acton _et al_., 1963) have demonstrated correlations between the degree of soil aggregation and the content of polysaccharides in various soil organic matter fractions. However, conflicting evidence has also been advanced (Harris _et al_., 1966). For example, Bloomfield (1956) found that polysaccharides of leaf extracts appeared capable of causing deflocculation of clays, permitting their translocation.

Most work on soil polysaccharides has involved the more persistent forms, believed to originate from (1) resistant or protected undecomposed plant residues, (2) microbial synthesis from non-carbohydrate substrates, and (3) microbial transformation of monomers released from decomposing

plant materials. Lowe (1978) noted that carbohydrates constitute approximately one tenth of soil organic matter, a proportion equal to or less than that found in stemflow and throughfall organic matter. Since these two sources contribute a small but mobile component of the total organic matter returned to the soil by forest vegetation, perhaps it is time for the role of this material in soil aggregation to be studied more thoroughly. It is also uncertain whether the soluble carbohydrates in the soil system, whatever their origin, are able to enter into the process of aggregate formation or are metabolized before that can occur.

Phenolic substances and uronic acids both have the ability to complex metal ions and have been implicated in the pedogenic mobilization of Fe and Al (Malcolm and McCracken, 1968; Davies, 1971). Such behaviour has also been demonstrated in artificial leachates of Populus spp. leaves (Dormaar, 1970, 1971). Since both of these organic substances were detected in the throughfall and stemflow samples at the study site in concentrations equal to or greater than those existing in the soil leachates, it appears that aboveground sources may have provided a significant proportion of the uronic acids and phenolics in the soil solution. Because Fe depletion was observed in the upper horizons of the soil at the study site, it is possible that these agents may have played a role. However, since no Fe was detected in the soil leachates, this may be unlikely and another mechanism, such as infrequent episodes of reducing conditions, may be responsible.

Duchaufour (1977) pointed out that the phenolic precursors of humic substances can have several fates, including: (1) precipitation of insoluble complexes formed by reaction with cations, (2) adsorption by inorganic colloidal components, (3) loss through biodegradation of side

chains and functional groups, and (4) condensation of aromatic nuclei. These processes are thought to be particularly active in mull Ah horizons in which the products of these reactions form the humic substances involved in clay-humic complexes.

Unlike the other soluble organic constituents, phenols were much less abundant in the soil leachates than in the throughfall and stemflow. It is pertinent that Davies (1971) suggested that in low rainfall climates, high base status soils will experience rapid mineralization of organic matter, including phenolic substances, thereby reducing the potential for Fe mobilization. Perhaps this explanation applies to Chernozemic soils, providing a buffering mechanism that slows the degrading influence of phenolic substances.

The question of elemental mobility in soils will be discussed next, since it governs the fate of nutrients entering the soil both in solution from outside and as released from decomposing organic matter. As shown by numerous studies (e.g. McColl, 1972; Johnson, 1975), the rate of cation transport in the soil is largely controlled by the anion content of the soil solution, since the negative charges of the exchange complex are immobile. In the temperate and tropical zones, the dominant anion is usually bicarbonate, formed by the dissociation of dissolved CO_2 produced by respiration in the soil. The hydrogen ions simultaneously released are available to replace exchangeable cations on colloid surfaces. In other environments, atmospheric sulphate pollution or the production of organic anions as a result of decomposition processes can reduce the importance of bicarbonate as a leaching agent. Clearly, though, the soil solution data from the Ellerslie site indicated that this is a bicarbonate-dominated system. The other anion determined, sulphate, apparently increased in

concentration with depth in the soil, but this observation should be treated with caution because of the limited number of analyses performed.

Differences in behaviour between elements have already been noted in this study; this discussion attempts to relate these differences to both the properties and dynamics of the soil system. As was already noted, K^+ concentrations peaked in the L-F leachate, reflecting the combined effects of leaching from both the living tree canopy and decomposing litter. In the absence of root uptake because of the lysimeter design, the consistent drop in K^+ concentration in the soil leachates with depth requires explanation. Since the clay mineralogical analyses indicated a significant content of mica and an undetermined but small amount of vermiculite, K-fixation may explain the drop in its concentration in the leachate between the H and Ah horizons. Despite the fact that about half of the K^+ in solution is lost during percolation through the Ah horizon, this cation occupies only about 2% of the exchange capacity, again suggesting the possibility of fixation. However, the drop in K^+ concentration in the H horizon is less easy to explain since the content of mineral material is quite low by comparison with the Ah horizon.

The other cations maintained more constant concentrations in the soil leachates, suggesting a higher degree of mobility than for K^+. This accords with the findings of Riekerk (1971) who observed Ca^{++} to be more mobile than K^+ in the mineral horizons of a forest soil. It is interesting that Na^+, although not abundant in throughfall or stemflow, did show an increase in its concentration in the Ah leachate. In fact, if the water chemistry data had included only those sampling dates when leachates from all three levels were obtained, the trend would have shown approximately a twofold increase in Na^+ concentration in the Ah leachate vs. that from

the L-F. This suggests some displacement of exchangeable Na^+, perhaps by K^+ which is higher in the lyotropic series (Gast, 1977); note that with depth in the soil, exchangeable K^+ decreased relative to Na^+. The abundance of Ca^{++} in the soil leachates presumably reflected its abundance in the leaf litter added to the soil and partially accounts for its abundance on the exchange complex. Ca^{++} will tend to dominate because divalent cations have a higher replacing power than monovalent cations. In addition, materials of high exchange capacity, such as the smectites, are abundant in the pedon, and favour adsorption of divalent cations (Wiklander, 1955).

As was observed in the earlier part of this chapter, mineralogical analysis suggested that there was only limited weathering occurring in the soil at the study site, a characteristic typical of Chernozemic soils. Since nutrient cycling processes exert such a decisive control over the dynamics of the biologically important elements in the soil, any conclusions about weathering processes would have to be based on the behaviour of other elements such as Si, Al, and Fe.

Although Fe was not detected in the lysimeter leachates, the pedon showed morphological and analytical evidence of iron redistribution. Hydromorphic features in the lower solum suggested that periodic reducing conditions had existed during the past. However, when a perched water table occurred at the site in the spring of 1979, *in situ* measurements of Eh and pH both in the Ah and Ckgj2 horizons indicated the persistence of oxidizing conditions. These findings were similar to those of Dormaar (1978) who observed that the brief period of saturation occurring during the spring defrosting process in Chernozemic soils was not sufficient to cause Fe reduction and migration. Therefore, the former drainage conditions

in the vicinity of the study area may have been poorer since more prolonged periods of saturation than occur today appear to be needed for Fe reduction and mobilization.

Aluminum in the soil solution would serve as a useful indicator of weathering, yet none was detected. This is not surprising, in view of the fact that the solubility of alumina at near-neutral pH conditions is almost nil (Mason, 1966). In addition, any small amounts of Al released by weathering would not necessarily be available to enter into the formation of new mineral species but could be tied up in non-crystalline complexes with organic acids, such as citric acid (Kwong and Huang, 1977).

Silicon concentrations in the Ah horizon leachates, although variable, showed an average value similar to those obtained by McKeague and Cline (1963) for soil extracts (0.9-15.0 mg l^{-1}). This finding coincides with that of Elgawhary and Lindsay (1972) who concluded that soluble Si in soils will occur at levels between those maintained by amorphous silica (51 mg l^{-1} at pH 8.0) and quartz (2.8 mg l^{-1}). However, the Si data in the present study have not been obtained from equilibrated solutions, so any detailed comparisons are pointless.

The foregoing illustrate the fact that the leachates sampled in this study represent only one part of the total soil solution. As demonstrated by Boudou et al. (1978), lysimeter waters differ in chemical properties from waters held at higher tensions in more intimate association with the mineral component of the soil. Therefore, it would be incorrect to relate the chemical data from the former to mineral stability fields and draw conclusions about the possible direction of mineral transformations. Tardy et al. (1973) made the same point in discussing granite weathering;

solutions in fine pores circulate more slowly than those in larger pores and therfore will have higher solute concentrations. Consequently, montmorillonite may be forming in the smaller voids at the same time as kaolinite is developing in the more freely drained pores. Analyses of only solutions from the latter will not give a true picture of the diversity of weathering environments in the soil.

5.3.5 Micromorphological Change

The two parent material cores emplaced in the surface horizons for two years showed the beginnings of pedogenic reorganization of fabrics. These changes were particularly marked in the core placed immediately beneath the litter layer. Prior to impregnation, the upper surface of that core had become quite irregular, with many rounded aggregates (1-3 mm diameter) being partially or completely detached (Plates 13-15). In addition, there appeared to have been some surface staining by organic matter, resulting in a slightly browner colour than at the bottom of the core. Litter materials had become partially mixed with the surface layer. In some cases, leaf petioles had been drawn into the larger voids, perhaps by faunal action. In thin section, it was clear that the structure of the upper half of the core had become more open. Planar voids seemed to have widened and there was a larger number of smaller (< 2 mm diameter) granoidic and fragmoidic units (Plate 14). Under magnification, the changes at the surface were quite evident, with newly-formed granic and granoidic units being intermingled with litter or humified material (Plate 15).

The core placed in the Ah horizon showed fewer changes, principally a more pronounced development of joint planes and some entry of mull-

Plate 13. Upper surface of parent material core placed beneath litter layer for 2 years (actual size). Note numerous rounded aggregates and partial incorporation of organic debris.

Plate 14. Granic unit (1 mm diameter) at surface of parent material core shown in Plate 13.

Plate 15. Granic unit incorporating 0.1 mm wide leaf fragments, formed at surface of parent material core shown in Plate 13.

granic units into the larger voids near the upper surface.

Several processes seem likely to have been responsible for these effects. The widening of planar voids could have resulted from a combination of frost action and the shrinking and swelling of the smectite-rich clays. The former possibly could be tested by retrieval of cores during the winter in order to examine them for development of segregated ice lenses. Biological activity was probably responsible for some of the incorporation of organic debris, particularly in the core placed below the litter layer. However, it is unclear how the emerging granic and granoidic units had formed near the surface of that core. It seems unlikely that faunal ingestion and excretion were responsible, since the material was essentially devoid of organic materials. Since aggregates of similar size and shape have been attributed to biological activity, this observation is interesting. Perhaps organic substances, such as polysaccharides, leached from the tree canopy and litter layer, and/or some unknown physical process operating near the soil surface were responsible for producing the rounded aggregates.

CHAPTER 6

SUMMARY AND SYNTHESIS

This final chapter brings together the principal findings of the study in order to assess the degree to which the central objective was met. Based on this assessment, some recommendations are made for directions to be pursued in further research.

One of the inherent limitations of a field-based study of soil dynamics is that the conditions prevailing during the investigation may be unrepresentative of those which have been typical during the pedon's evolution. However, this difficulty should not be permitted to discourage such studies. Observations of and experimentation with soil-forming processes are necessary adjuncts to traditional approaches in soil genesis research.

The most obvious example of this problem with short-term studies of soil dynamics comes in the effects of year-to-year variations in climate. However, such fluctuations provide an opportunity to observe any related changes in the behaviour of the soil system. For example, the June to August precipitation in 1979 was 20% lower than during the previous year, with summer rainfall during the two years falling on both sides of the long-term mean. Another significant and related difference was in the springtime moisture status; a brief period of saturation occurred at the site in 1979, but not in 1978 or 1980. Despite these variations in moisture status, at no time was there evidence for iron reduction and mobilization. This finding was interpreted to indicate that the hydromorphic features in the pedon are relict and were formed under more poorly drained conditions in the past. Given the Land Survey evidence for exten-

sive areas of slough and marsh in the Lake Edmonton plain prior to land clearance and settlement, it seems likely that soil drainage conditions have changed during the past century. In that sense, then, it appears that some of the pedon's morphology can be attributed to conditions which no longer exist in the present environment.

Biological activity plays a major role in pedogenesis at the site, both as inferred from the morphological and analytical properties and as observed through the process studies. The principal findings were as follows.

(1) Considering only the litterfall from trees and shrubs, the supply of organic matter from those sources lies within the upper range found in similar deciduous forests in the cool temperate zone.

(2) Understorey species, particularly Cornus stolonifera, made a major contribution to the litterfall and appeared to play an important role in base cycling, notably of calcium. This resulted from both a rapid rate of decomposition and a high initial content of calcium in the litter from this species. Faunal and microbial activity indicated that the litter of this species was an attractive substrate.

(3) Fecal materials produced as a result of faunal attack on leaf residues comprised much of the organic horizons. In addition, some faunal mixing of mineral and organic material appeared to have taken place at the lower boundary of the H horizon.

(4) Despite the contrast in growing season precipitation totals between the two years of the study, both periods showed the same pattern of progressive moisture depletion from the solum during the summer months, principally because of the transpiration demands of the forest vegetation. The implication is that leaching events in this environment are restricted

to the spring thaw period and occasional heavy rainfall episodes during the summer and fall. Therefore, it seems incorrect to assume that a substitution of forest for grassland vegetation necessarily enhances the potential for leaching. In addition, the interception and moisture storage effects of the soil organic horizons also tend to limit the degree of leaching. However, periodic forest fires could remove part or all of the surface organic matter accumulation, thereby limiting this barrier to leaching processes.

(5) Within the Ah horizons, the granular soil structurre resembled that which is typical of the mull humus form. While a faunal origin is possible, it was significant that aggregates similar in size and shape to those in the Ah horizons were observed to form after two years in parent material cores placed near the soil surface. These structures may have resulted from physical processes and/or the aggregating effects of soluble organics in the soil solution.

(6) While the high base status, particularly of calcium, seemed related to the considerable nutrient return in litterfall, an additional factor was the effect of forest canopy leaching on soil solution chemistry. Considerable amounts of K^+ were likely returned to the soil by this pathway. Canopy leaching also seemed to be important as a source of soluble organics for the soil solution. Uronic acids and saccharides from this source may play a role in aggregate formation in the solum. Phenolic substances were also released in throughfall via canopy leaching and appeared to diminish in abundance in the soil solution, with several mechanisms possibly being responsible, including adsorption by inorganic colloids and polymerization.

(7) Soil organic matter contents were comparable to those in

other Black Chernozemic soils. Qualitatively, the clay fraction showed a high degree of clay-organic complexation, particularly in the <0.2 μm fraction. Although the interpretation may be open to controversy, infrared spectra of humic acids from both organic and mineral horizons differed from those typical of Chernozemic soils. However, these differences may be attributed to the effects of forest vegetation or hydromorphic conditions.

The high base status and near-neutral pH are indicators of a mild weathering environment, an interpretation borne out by the lack of clay transformations. Biological cycling of certain mineral-forming elements (e.g. Ca) made it difficult to assess weathering processes on the basis of soil solution data. Silicon concentrations in the soil solution were also difficult to relate to weathering processes because of the dominance of dilution effects controlled by precipitation volume.

The foregoing evidence and interpretations are consistent with the hypothesis that the pedon owes many of its features to genetic processes operating in the present environment. In particular, the evidence of direct and indirect biological effects on such aspects as structure development, organic matter incorporation and transformation, and base status appear to support this hypothesis.

The picture of Holocene environmental change in central Alberta that has emerged from recent studies suggests that it may not be correct to envision the soils of the northern part of the Black soil zone as relics of a prolonged Altithermal grassland invasion. Pollen studies, indicating apparent stability in vegetation zone boundaries for the past 2000-3000 years, in addition to the paucity of grass-derived plant opal in soils of that region, suggest that forested conditions

have been established in the Edmonton area for several centuries. Because of this evidence, the persistence of Black Chernozemic soils displaying only limited eluviation seems to contradict conventional ideas about the rapidity of soil change resulting from forest invasion of grassland.

While these conclusions about soil genesis are necessarily tentative, and appear to contradict the generally accepted soil biosequence in western Canada, it is useful to return to some of the points raised in Chapter Two. Although Canadian soil classification systems have often borrowed terminology from Russian and European pedology, correlations between classifications in the two continents are often hampered by differing definitions and genetic concepts. Many factors may be responsible for such problems: incomplete or faulty translation, differing experiences and attitudes on the part of pedologists, and inherent differences between soils originally thought to be similar in North America and Europe.

Of particular relevance to this study is the fact that Canadian pedologists have given a much broader definition to the Chernozemic soils than have their European counterparts. As pointed out earlier, the sequence of genetic soil types recognized in the U.S.S.R. along the Boreal Forest to steppe biosequence is Sod-Podzolic, Gray Forest, and Chernozem. In Canada, the analogous sequence consists only of the Chernozemic and Luvisolic soils, with there being no explicitly recognized counterpart of the Gray Forest soils. Since the latter type is characterized by a prominent Ah horizon and some degree of eluviation, they are a distinct intermediate type. While Russian pedologists consider these to be the typical forest soil occupying the transition between the steppes and

the Boreal Forest, Canadian pedologists seem to feel that the persistence of any substantial Ah horizon is incompatible with pedogenesis in the mid-continental forest environment. This latter view ignores the existence of Melanic Brunisols with well-developed mull Ah horizons which have formed under deciduous forests in the Great Lakes region. Moreover, Orthic Sombric Brunisols with well-structured dark Ah or Ahe horizons (approximately 10 to 12 cm thick) and showing evidence of strong earthworm activity, often occur on eastern Vancouver Island under coniferous forests and on acidic parent materials. Since genetic concepts are explicit or implicit in the Canadian soil classification system, it would be better to adopt a more restricted definition for the Chernozemic soils that is in keeping with the original usage of the term.

It might also be fruitful to consider an alternate genetic pathway, that of the Meadow Chernozems, as being applicable to certain of the Black soils of western Canada. The apparent hydromorphic influence on pedon development at the study site may be a product of such a pathway. The recognition, at a high taxonomic level, of the Gray Forest soils or some clearly identified equivalent would provide a useful grouping for the current Dark Gray Chernozemics, as well as some of the Black Chernozemics, notably the Eluviated subgroups. This change would allow these soils to be viewed as a distinct and stable product of their forest environment, rather than as transitory phases in an assumed degradation sequence. As matters stand now, Canadian soil taxonomy has borrowed only some of the concepts and terms involved in the continental soil biosequence originally identified in Europe. The suggested modifications would correct these omissions and perhaps bring out some useful contro-

versy as to the genesis of the Black Chernozemic soils.

While several lines of future research are suggested by the findings of this study, there are three immediate recommendations. First, it would be useful to conduct a matching study, using similar techniques, to observe the dynamics of a Black Chernozemic soil system at a grassland site. Since many of the observations made during this study pointed out the important role of the forest cover in, for example, soil moisture depletion and chemical modification of precipitation, it would be useful to have comparative data from a grassland environment. Second, the experiment involving the observation of fabric changes in sample cores of parent material should be continued at this site and as part of the companion study suggested above. Additional treatments could include placement of cores adjacent to tree trunks, so as to observe the effects of stemflow, and the exclusion of litter from the core surface, so that only effects of physical processes could be monitored. Finally, the limited opal phytolith data presented in this study pointed out the considerable variation in soil opal content across the Black Chernozemic zone. Since this information has significance for both environmental history and soil genesis research, there is an obvious need for systematic regional studies of the amounts and types of plant opal in Alberta soils.

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APPENDIX

CANADIAN DEFINITIONS OF CHERNOZEMIC SOILS

The Chernozemic A horizon, as defined in The Canadian System of Soil Classification (Canada Soil Survey Committee, 1978), has the following properties:

1. It is at least 10 cm thick.
2. Its color value is darker than 5.5 dry and 3.5 moist, and its chroma is lower than 3.5 moist.
3. Its color value is at least one Munsell unit darker than that of the IC horizon.
4. In soils disturbed by cultivation or other means the Ap horizon is thick and dark enough to provide 15 cm of surface material that meets the color criteria stated in 2 and 3 above.
5. Its organic C content is 1-17% and its C:N ratio is less than 17.
6. Characteristically it has sufficiently good structure so that it is neither massive and hard nor single grained when dry.
7. Its base saturation (neutral salt) is more that 80% and Ca is the dominant exchangeable cation.
8. A chernozemic A horizon is restricted to soils having a mean annual soil temperature of 0°C or higher and a soil moisture regime subclass drier than humid.

The Black Chernozemic Great Group has a Chernozemic A horizon with a colour value darker than 3.5 dry and a chroma usually 1.5 or less, moist.